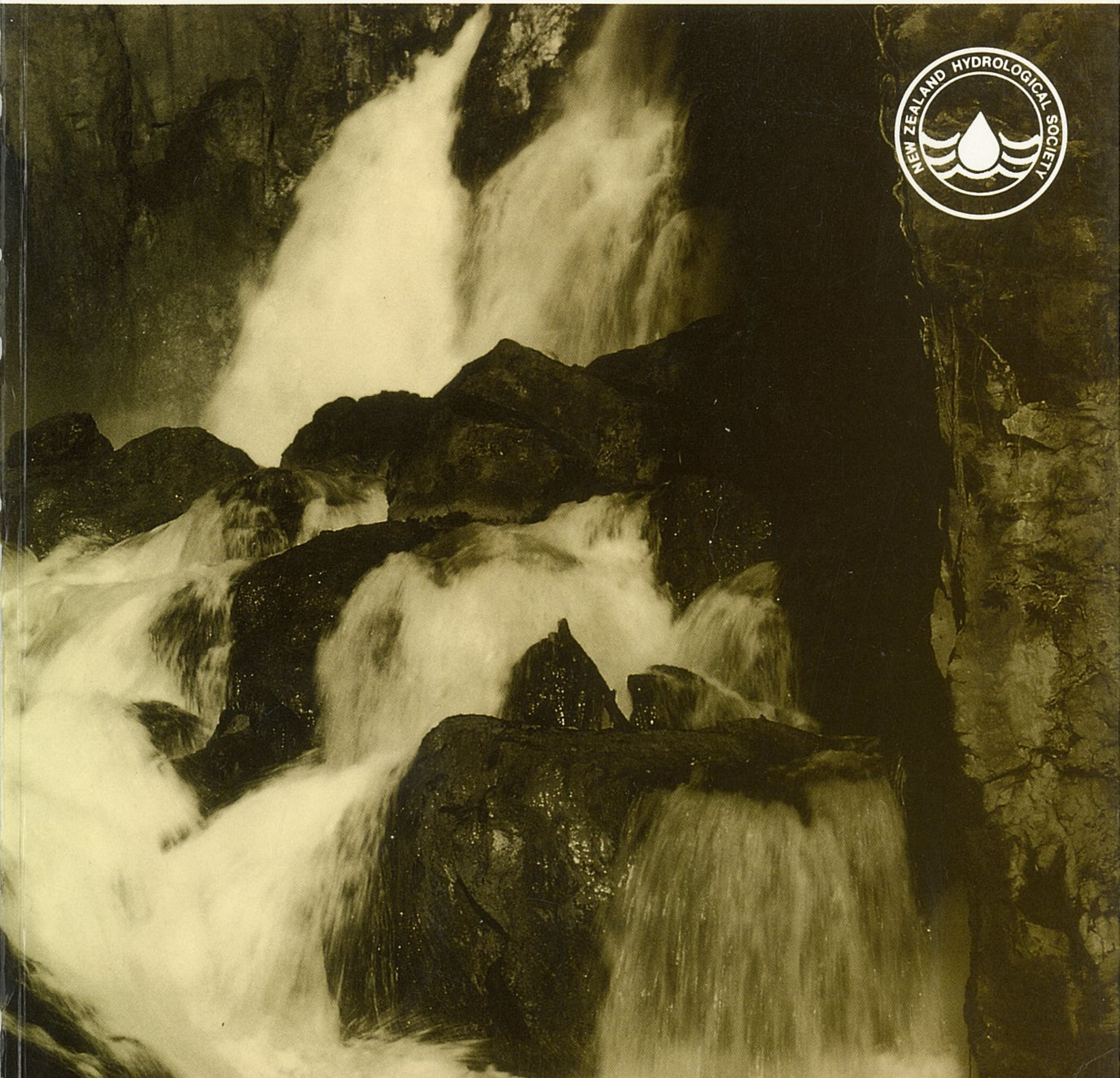
Waters of New Zealand



edited for the New Zealand Hydrological Society by **M Paul Mosley**

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First published 1992

Cover photograph by David E. Harding

ISBN 0-473-01667-2

Published by New Zealand Hydrological Society Inc
Post Office Box 12-300, Wellington North, New Zealand

Printed by
The Caxton Press, Christchurch

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Preface

The aim of the New Zealand Hydrological Society is to promote the science of hydrology - the study of water - in New Zealand. We started work on this book in the thirtieth year of existence of the Society. We wished firstly to commemorate the contribution to New Zealand's quality of life which has been made by hydrology and hydrologists. More than that, however, we wanted to provide a resource for students of hydrology, particularly at university and senior school level. Knowledgeable, skilled people are a country's greatest asset; by making this book available we hope to increase the level of knowledge about the waters of New Zealand.

The authors of the chapters are amongst New Zealand's leading hydrological scientists and practitioners. Their knowledge covers the entire scope of the hydrological cycle on the land, from rainfall and snowfall through to deep groundwater. They provide a New Zealand perspective on hydrology and water resource management which gives a much needed supplement to the texts - mostly British or North American - normally used by students. Certainly, the physical and chemical principles which govern the behaviour of water are the same in New Zealand as elsewhere in the world. Nevertheless, water is a very distinctive element of the life and landscape in New Zealand, and the ways we use and manage it are often characteristically "New Zealand". This book summarises much of the work of New Zealand hydrologists over the years, particularly that work described in the Society's own **Journal of Hydrology (NZ).**

One of the most characteristic aspects of water management in New Zealand is almost absent from this volume, however. This is the perspectives on water management taken by Maori. Maori concerns relating to safeguarding water quality, aquatic ecosystems, and food resources have amplified and accelerated recent efforts to manage water more carefully, rather than initiated them. Nevertheless, significant Maori input to water resource management is desired by the community,and indeed is mandated in law. The Society, then, is particularly conscious of the almost complete lack of Maori hydrologists, and the difficulty of finding Maori who are able or willing to present Maori perspectives in a format such as this. Hydrology is firmly founded in the western scientific method. New Zealand's hydrologists and water resource managers are slowly learning to combine their own approach with those of other sections of the community.

Hydrology is a somewhat schizophrenic discipline, in some respects purely scientific, and in others very much applied. Recently, hydrologists have been questioning the real nature of hydrology: is it simply a part of engineering, or does it have a scientific basis which qualifies it to be considered as part of the natural sciences as a whole? New Zealanders are great pragmatists, and the content of this book may well persuade our readers that the answer is "yes" to both questions, and it works. Much hydrology in New Zealand is now supported by the Public Good Science Fund, administered by the Foundation for Research Science & Technology. At present, we are not certain whether the Foundation wishes to place its greatest emphasis on the scientific excellence of the work it supports, or on its ability to solve problems. Again, the answer may well be "yes" to both, an answer which this Society would happily endorse.

Hydrology and water resources management are experiencing a time of change and uncertainty not experienced before in New Zealand. Successive pieces of legislation have brought new directions over the last half century: the Soil Conservation and Rivers Control Act 1941; the Water and Soil Conservation Act 1967 and its amendments (particularly the 1981 "Wild and scenic rivers" amendment); and most recently the Resource Management Act 1991. Each has reflected a new emphasis: erosion, sedimentation, and flooding in the 1930s; water quality in the 1960s; the growing importance of recreation and amenity in the late 1970s; recognition of the need for resource management to be sustainable in the 1980s. In each phase, hydrologists and water resource managers have had to turn their sights onto new issues, sometimes following and sometimes leading the community at large.

It is interesting to speculate about the emphases in hydrology in the first decade of the next century. One might choose from a range of issues that are already evolving: impacts of climatic change, the effects of toxic waste, critical water shortages and the need for new technology to conserve water, the wholesale use of economic instruments as management tools. With growing competition for increasingly scarce water resources, we can be certain that hydrologists will have a major role to play in maintaining the quality of life in New Zealand.

It will also be interesting to see how well the recent sweeping reorganisations of national and regional government will serve the community. In particular, the Society hopes that restructuring of regional and local government and establishment of the Crown Research Institutes will provide an environment in which hydrology can truly be applied for the good of all New Zealanders. Certainly, that is our aim. We hope that this volume will contribute to it.

M Paul Mosley

Wellington, New Zealand
August 1992

Acknowledgements

We must thank, above all, our employers for the opportunity to participate in this venture. Much of the work on these chapters has probably been in the authors' own time, but they have assembled their knowledge "on the job", and for that we are grateful. In addition, however, the resources of our employers, particularly for drafting and word processing, have frequently been drawn on; we hope that the end product will be sufficient recompense. Individual chapter authors make specific acknowledgement of the work of reviewers, permission to reproduce figures, and so forth.

Much hydrological research in New Zealand is now supported by the Public Good Science Fund administered by the Foundation for Research, Science & Technology. Our thanks must go in particular to the Foundation for the part it has played - perhaps indirectly and even unwittingly - in enabling this book to be produced.

This last year has not been a good time to take on extra tasks. As editor, then, I wish personally to acknowledge the hard work of the chapter authors themselves, the various reviewers, and the three people who have played a particular role in various phases of production: Pauline Freestone, Marisa Lovato, and Eileen McSaveney.

1
Introduction: Hydrology in New Zealand

John Waugh

New Zealand is a small country with only 3.4 million people and a land area of 270,000 km^2, but on a world scale it is well-endowed with fresh water. Prevailing westerly winds blowing across the ocean bring abundant precipitation, especially to the western sides of both main islands. Twelve metres or more of rainfall each year deluge parts of the western Southern Alps of the South Island (Griffiths & McSaveney, 1983), but rainfall rapidly decreases as one moves east. For example, mean rainfall is 6240 mm/yr at Milford Sound on the West Coast of the South Island, but decreases to 340mm/yr at Alexandra in Central Otago.

While the total water resource of New Zealand rivers is about 300 km^3/yr, less than 2 km^3/yr (0.64%) is used for domestic water supply or for agriculture and industry (Department of Statistics, 1989). Dwarfing this water use is the 100 km^3/yr of flow used in hydro-electric generation, but this water may be used several times as it passes through successive dams on its way down rivers like the Waikato or Waitaki. Eighteen hydro-electric power stations are located on these two rivers alone.

Although abundant, New Zealand's water resources are not well distributed. The eastern areas of both islands normally have dry summers and suffer seasonal soil moisture deficits which are a major constraint on horticultural development. Surface water and groundwater have been widely exploited for irrigation. Irrigation is used on about 234,000 ha of agricultural land (Department of Statistics, 1989), and most large irrigation schemes are in the drier eastern areas, such as Canterbury and Otago. In Canterbury, the Rangitata Diversion Race (66 km long) takes up to 30 m^3/s of flow from the Rangitata River and about 4 m^3/s from the South Ashburton River, to supply three irrigation schemes, Mayfield-Hinds (35,000 ha), Valetta (6,900 ha) and Ashburton-Lyndhurst (26,000 ha). Surplus water generates electricity at Highbank Power Station before discharging to the Rakaia River (Walsh & Scarf, 1980).

In the South Island hydro-electric catchments, mainly the Clutha and Waitaki, much of the precipitation falls in summer. Electricity demand, however, peaks in winter, thus presenting electricity generation agencies with a major water storage problem.

Floods are a common hazard in New Zealand, as many of the larger towns and cities are built on river floodplains. In spite of extensive river control works, severe floods still cause major damage and disruption to communications and farming. For example, a severe flood in South Canterbury on 13 March 1986 caused $60 million of damage and led to some 2,000 people being evacuated as part of the Regional Civil Defence emergency, (SCCB, 1987).

Most New Zealanders, with perhaps the exception of farmers, take water resources for granted; perhaps a reflection of the abundance of water in New Zealand. Public awareness of environmental issues is high and any pollution of water resources generally draws a strong public reaction. Large numbers of people participate in water-based recreation such as boating and fishing, and rivers, lakes and seashores are popular for summer holidays. In recent years there has been a rapid growth of outdoor "adventure" recreation activities involving jet-boating, rafting and canoeing. These activities are important to the developing tourism industry.

Since the Water & Soil Conservation Act was passed in 1967 there has been a growing desire by the public to safeguard water bodies. This led to an amendment in 1981 which allowed "wild and scenic rivers and lakes" to be protected by Water Conservation Orders. Over the last two decades there has also been major upgrading of waste disposal, with a resulting improvement of water quality.

The quality of New Zealand water is generally high, but in some catchments it has been affected by effluents from urban areas and facilities such as dairy and wood processing plants, by runoff enriched by fertiliser and animal wastes from agricultural areas, and by sediment introduced by accelerated erosion. Runoff from urban areas and highways has also been implicated in pollution of some rivers and estuaries.

New Zealand has, over the years, developed a comprehensive system of organisations and legislation as a basis for water resource management and planning. The most significant recent development is the Resource Management Act of 1991, which replaces a patchwork of almost 80 earlier acts and regulations, and amends over 50 others.

Hydrology in New Zealand : a Brief History

New Zealand has had a long involvement in hydrological investigations. The earliest water level recorders were imported by the Public Works Department in 1898.

Most of the early hydrological work in New Zealand was carried out to investigate the potential for hydro-electric power generation. By December 1912 a monthly chart recorder was installed on the Wairua River, Northland, above the Wairua Falls to investigate a hydro-electric power scheme. P.S. Hay (1904), Superintending Engineer of the Public Works Department, records flow measurements of the Wairua River and the Waikato Rivers in 1903, and current meter gaugings were carried out on the Wairua between 1911 and 1913. The longest continuous hydrological records available in New Zealand are for Lake Taupo and Lake Rotoiti, North Island, where daily observations commenced in 1905 (H. Freestone, pers. comm.). In the South Island, records of water-level were collected at Lake Wakatipu during 1901-1903. In the 1920s and 1930s hydro-electric power investigations continued, with water-level recorders being established on major lakes, notably Wakatipu (1926), Wanaka (1929), Hawea (1929), Taupo (1905), Waikaremoana (1929) and Coleridge (1929). The data from these early stations are available in computer archives.

Irrigation investigations stimulated new hydrological investigations in Canterbury in the late 1930s, with water-level recorder stations being established on rivers such as the Rangitata (1936), Opihi (1935), Opuha (1936), South Ashburton (1936), North Ashburton (1938) and Hurunui (1938). Some of these stations (Opihi and Opuha) have continued to operate up to the present day, and thus have among the longest river flow records (non lake-fed) in New Zealand. Unfortunately, World War II disrupted much of this early hydrological work, and at many sites insufficient flow gauging was carried out.

The Soil Conservation and Rivers Control Act (1941) and the subsequent establishment of Catchment Boards gave a major boost to hydrological work in New Zealand. To assist the Soil Conservation and Rivers Control Council, in 1949 the Ministry of Works established "Hydraulic Survey" parties, initially located at Hamilton, Palmerston North and Blenheim, employing such noted personalities as "Hoppy" Hopkins, C. Toebes, W. B. Morrissey and E. J. Speight.

New Zealand participated in UNESCO's International Hydrological Decade (IHD) from 1964 to 1975, establishing 53 representative basins distributed throughout the 90 hydrological regions defined by Toebes and Neef (1962), and Toebes and Palmer (1969). Small experimental basins to examine the effect of land-use change were also established, and research catchments such as Moutere, Makara, Taita, Otutira, Purukohukohu and Puketurua date from this period. A few are still operating today; while others have been closed but their data are still being used. The representative basins programme greatly widened New Zealand's hydrological network to sample a range of catchment sizes and to monitor catchments other than the larger rivers already monitored for power, flood-control and irrigation purposes.

In the late 1960s, with the rapid growth of the hydrological network, manual data processing became an impossible task. New Zealand adopted punched-tape water-level recorders and computer-based data-processing. S. M. Thompson played a pivotal role in the development of Tideda, a time-dependent data processing, storage, analysis and modelling system. It is now a sophisticated computer package which is used extensively throughout New Zealand and several other countries. C. Toebes recognised the importance of modern equipment, sound field methods and computer - based data processing. He played a leading role in hydrology during the 1960s and early 1970s, and did much to firmly establish the science of hydrology in New Zealand.

In the 1970s Water and Soil research activity was consolidated, and three research centres were established at Hamilton (Water Quality), Aokautere near Palmerston North (Land Resources) and Christchurch (Hydrology). The Water Resources Survey emerged as a separate entity and all of these groups were moved to the Department of Scientific and Industrial Research in 1988 and then to the National Institute of Water and Atmospheric Research in 1992.

During the 1980s there was rapid growth of data collection by Catchment Boards, which merged into Regional Councils in 1989. Radio-telemetry of hydrological data became widely used during the 1980s for flood-warning, water-resource management and basic data collection. At the same time, the advent of cheap computers enabled data processing and archiving to be decentralised. The creation of local or regional data-bases, however, created problems in maintaining a national hydrological archive. Quality assurance concepts (Mosley and McKerchar, 1989) were accepted to ensure that data entering archives met specified standards. The existence of a single national hydrological archive of high quality data is a major asset, which is particularly valuable for nationwide studies such as flood frequency (McKerchar & Pearson, 1989). Another important hydrological database is the "Power" database now operated by Works Project Services, formerly Ministry of Works and Development Power Division. This database contains inflow records for most of New Zealand's lakes, plus flow data relating to electric power stations in New Zealand.

The growth of New Zealand's hydrological network is reflected by the data available for flood studies. In 1982 Beable & McKerchar, using data up to 1978, had 160 stations available for analysis. By 1989 there were 343 suitable stations available. Of these, 275 stations had records of 10 or more years, with an average record length of 21 years (McKerchar & Pearson, 1989). In 1959 New Zealand had 109 permanent flow stations operating, but by 1989 the network had increased to 915 permanent water-level recording stations (M Duncan, pers. comm and Acheson, 1968).

The Hydrological Cycle

The world's oceans are huge reservoirs from which all water originates and to which all water returns. Not all water, however, completes the entire hydrological cycle at all times. There are loops in the system, for instance when water evaporates from land or saturated forest canopy and then returns to the land as precipitation.

In the complete hydrological cycle, (Figure 1.1), water evaporates from the oceans and the resulting moist air moves inland, condenses to form clouds, and moisture is released to fall to the earth as precipitation. In New Zealand precipitation occurs in all its forms, as rain, hail, fog, snow, rime and hoar frost.

Snow and Ice

Perennial snow and ice form a small but important part of the hydrological cycle, especially in the South Island of New Zealand. Anderton (1973) identified some 527 glaciers in New Zealand, with a total surface area of 810±40 km^2 and an es-

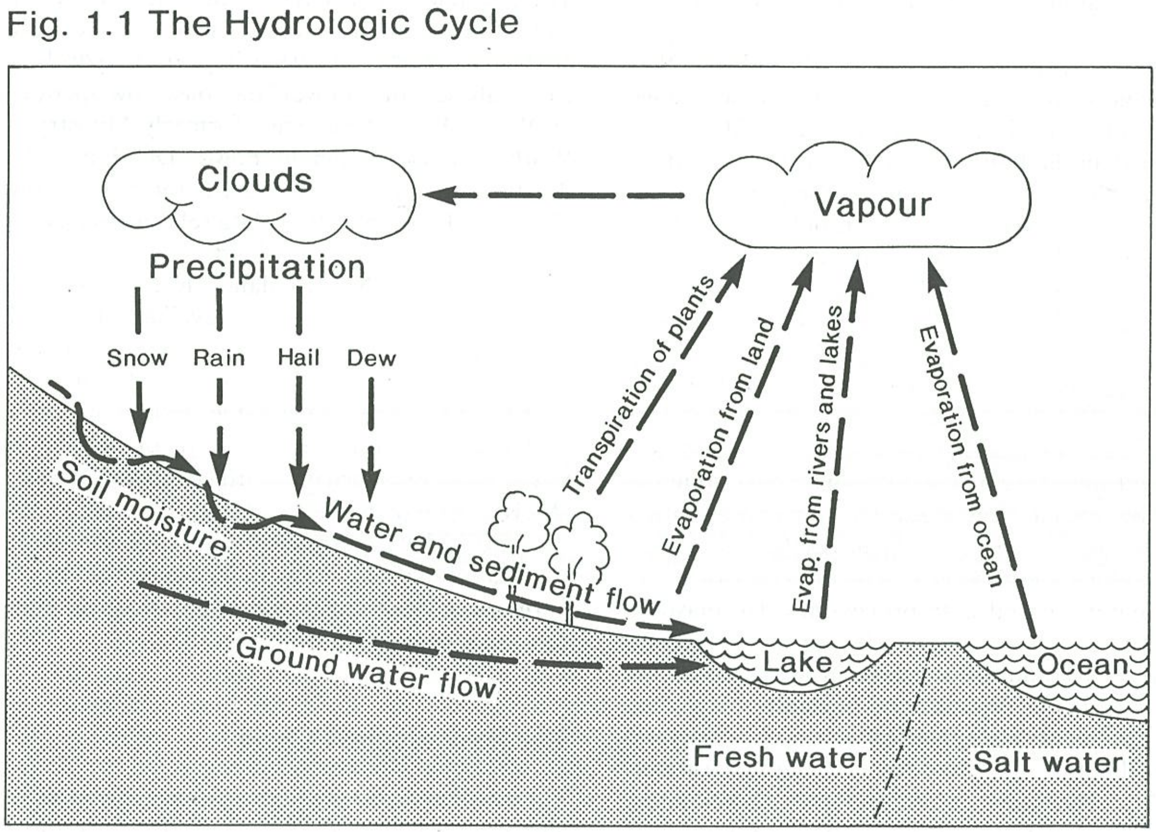

Figure 1.1 The Hydrological Cycle

timated volume of around 50 km^3. During this century New Zealand's glaciers have rapidly decreased in volume (down-wasting) with a consequent retreat of the glacier snouts. On the eastern side of the South Island main divide new, quite large glacial lakes have formed in valleys such as the Tasman, Hooker, Hopkins and Godley, since the mid-1970s. Anderton also showed that, between January and March, glacier meltwater contributes about ten percent of the inflows to Lake Pukaki. This estimate does not include meltwater from the much more widespread and variable seasonal snow pack. Snow and ice help to regulate streamflow by tending to increase summer flows, particularly low flows during dry summers, and by decreasing winter flows relative to glacier-free basins (Anderton, 1973).

Evaporation

Another aspect of the hydrological cycle which has received considerable attention in New Zealand is interception of precipitation by plants and its evaporation back into the atmosphere. This is one of the "loops" in the main hydrological cycle.

Early estimates of evaporation rates for New Zealand were given by Finkelstein in his 1973 paper on tank evaporation. He reported annual average evaporation in the range 650-850 mm in the North Island and 700-950 mm in the South Island. Toebes (1972), in his water balance study, suggested the evaporation component for the country was around 600 mm.

A study by Freestone (1981) of 59 NZ representative basins showed evapotranspiration losses, plus deep groundwater losses, if any, to average 720mm for the North Island and 540mm for the South Island. Freestone suggested that the low evapotranspiration loss for the South Island reflects the under-recording of rainfall in wet mountain catchments.

Interception, that is the trapping of precipitation on the surfaces of trees, tussock, tall grass and pasture, and its subsequent evaporation back into the atmosphere, has been intensively studied in New Zealand, especially interception by forest canopies.

Fahey (1964) reported on interception and throughfall in a stand of radiata pine near Dunedin. In more recent years there has been considerable work in both forests, particularly by Forest Research Institute at the Maimai catchments, Rowe (1979), Pearce & Rowe (1979) and Pearce, Rowe and O'Loughlin (1982), and in tussock grasslands (Mark and Rowley, 1976). Forests in New Zealand generally intercept between 20 and 35 percent of total rainfall, although Fahey's early work in radiata pine near Dunedin reported that 49% of rainfall was intercepted.

Infiltration and Soil Moisture

Infiltration is the movement of rainfall into and through the soil mantle. It received some early attention in New Zealand (Toebes, 1962), and numerous attempts were made to carry out field measurements of infiltration capacity for various New Zealand soils or catchments (Blake et al, 1968). Understanding of infiltration capacity of soils is important in relation to agriculture and irrigation. It is also important in attempting to construct mathematical models of flood runoff. Watt (1979) summarises soil moisture investigations in New Zealand and their relation to water balance studies. Partly because of the difficulty of extrapolating from point measurements to larger soil areas or whole catchments, detailed measurements of both infiltration and soil moisture largely gave way to catchment monitoring of flow and rainfall during the 1970s. Nevertheless, where detailed measurements of soil moisture are available within a larger monitored catchment, as at Moutere (Duncan, 1991) and Purukohukohu, they give considerable insight into the changes in the moisture budget of the catchment and are a valuable adjunct to the wider catchment monitoring.

Groundwater

As in other parts of the world, groundwater has long been exploited for household and city water supplies. From earliest settlement, towns such as Christchurch, Petone and Hastings have relied on groundwater to supply high-quality potable water. Grant (1965) reported on the ground waters of the Heretaunga Plains and identified the Ngaruroro River as a major recharge source. In the last two decades (1970s and 1980s), as pressure on surface water resources has intensified, there has been much greater use of groundwater and extensive development of irrigation from pumped wells, especially in Canterbury.

Groundwater in New Zealand is an important part of the hydrological cycle with water being added to aquifers from rainfall on the land surface e.g. Canterbury Plains and by recharge from rivers (Grant, 1965). The groundwater in gravel aquifers of the Canterbury Plains represents a very large storage of water. Groundwater studies and hydrogeology developed rapidly during the 1970s and 1980s and model studies of aquifers made major advances of understanding possible. Reports such as Hunt and Wilson's (1974) work on aquifers in Canterbury, and Donaldson's (1974) model study of the underground waters of the Lower Hutt valley provided tools for the management of important groundwater resources. Ministry of Works & Development (1977) carried out a major investigation of contaminant dispersion in the Heretaunga Plains aquifer. Computers and modern well-logging equipment are essential tools for the modern groundwater hydrologist.

Runoff and Streamflow

Surface water runoff is the part of the hydrological cycle where water, either directly from rainfall or via soil and underground aquifer storage, collects into streams, lakes and rivers and flows back to the oceans, (Figure 1.1). With climate change, the amount of moisture in particular parts of the system may change, for example, less water may be stored as glacial ice in response to increasing temperature.

Streamflow measurements are carried out throughout New Zealand by field staff of the Water Resources Survey and Regional Councils. Data are stored in national and regional archives using the Tideda system and numerous forms of output or analysis can be produced to meet requests for information.

Because of the commitment to make data available in electronic form, New Zealand does not now publish an annual water "year book", the last "Hydrology Annual" being published for 1967. However, an Index to Hydrological Recording Sites in New Zealand (Walter, 1989) has been produced every two years, to provide a guide to sites for which data are available.

Streamflow and Land-Use

The changing use of the New Zealand landscape has often had a measurable effect on the nation's rivers and streams. Starting with the IHD (1965-74) there have been numerous studies of the impact of land-use change on streamflow, water quality and sediment yield. Both Water and Soil Division of Ministry of Works and Development and Forest Research Institute staff were actively engaged in research projects based on monitored catchments throughout New Zealand. Paired-catchment studies have generally been valuable particularly when complemented by detailed "process" studies within the monitored catchments.

Water Quality and Sediment

Early hydrological work in New Zealand concentrated on engineering applications such as hydro-electric power investigations, irrigation and flood control. Because of this emphasis, streamflow measurements and water-level recording were the dominant activities until around 1970. On some rivers suspended sediment data were collected and sediment rating curves were established. A major use for these data was to determine sedimentation rates in hydro-electric power lakes.

Increasing public awareness of environmental issues such as river pollution saw measurement of water quality parameters commence in earnest during the 1970s. Major studies of the nutrient inputs to Lake Taupo (Schouten, et al 1981) and Lake Rotorua (Rutherford, 1984) were carried out and by 1976 investigations had extended to the quality of groundwater.

The first really comprehensive water quality and biological monitoring programme in New Zealand, "The 100 Rivers Project", was carried out in the late 1980s and reported on in 1990. It characterised, classified, and modelled New Zealand's rivers according to their hydrological, water quality, and biological properties. The results are detailed in eight research papers (Biggs et al 1990)

Results of the project show that New Zealand waters were of high quality, with high dissolved oxygen levels, pH generally in the range of 7.0 - 8.5, and low turbidity under baseflow conditions. New Zealand rivers also have much lower concentrations of plant nutrients and ions than the world average for freshwaters.

Water temperatures were sampled as part of the 100 Rivers Project and earlier records have been summarised by Mosley (1982). Most New Zealand rivers are temperate, cool-water systems with average winter temperatures rarely below 6°C, and average summer temperatures rarely above 22°C, reflecting the country's latitude, maritime climate, and the elevation of the sites.

The generally high quality of New Zealand river waters reflects the relatively small population and low level of industrial development of many rivers chosen for the 100 Rivers Project. The sites are well distributed throughout the country, in the upper parts of most catchments, and with high specific discharges (numerous floods and freshes) there is plenty of water to flush the river channels.

Almost all the rivers sampled in the 100 Rivers Project met existing water quality guidelines and standards for drinking water and freshwater fisheries.

Sediment

A sediment gauging programme was established by the Ministry of Works in 1959 to estimate the erosion rate of upstream catchments, and to provide background sediment data for the whole country. Suspended sediment gaugings have been carried out on many rivers by hydrological field parties, the samples analyzed and sediment rating curves prepared using standard methods derived from US Geological Survey practice, Hopkins (1959 a, b, c).

Thompson and Adams (1979) reported that within New Zealand there are catchments of 1,000 km^2 eroding at rates from 40/tonnes per square kilometre per year up to 8,000 (Jones & Howie, 1970), by world standards a wide range for a small land area.

A major study by Jowett and Hicks (1981) of bedload sediment transport was carried out on the Shotover River, a tributary of the Clutha River. It showed that material coarser than 0.3mm was transported as bedload, forming 14% of the total sediment load in the Shotover River. 80% of the sediment is trapped in Lake Roxburgh on the Clutha River. Griffiths (1981 and 1982) provided useful means of estimating suspended sediment yields of catchments in the North and South Islands.

The colour and clarity of waters have a strong influence on their aesthetic appeal and suitability for recreational use. The colour and clarity of New Zealand's rivers were reported on by Davies-Colley and Close (1990) for the first time, as part of the "100 Rivers Project". Mutual relationships were examined between colour (hue), visual clarity, turbidity, suspended solids etc. Visual clarity of water was shown to be strongly related to turbidity. Optically pure water in New Zealand has a blue hue, and "clear" rivers tend to be blue-green in hue. Water has an orange hue where there are high concentrations of organic material, and turbid waters are more typically yellow-coloured.

New Zealand's Water Resources

The estimated total discharge of New Zealand rivers is about 300 km^3/yr. About 50 km^3 is stored in perennial snow and glacier ice. Biggs et al (1990) compared the sample of rivers in the "100 Rivers Project" with rivers in the continental USA and conclude that the flow variability and specific yield (flow per unit area) of New Zealand rivers were very high. "These are fundamental properties of New Zealand river ecosystems which will have a major bearing on why our rivers are of such high chemical quality compared with elsewhere in the world, and why they are extensively dominated by "clean-water" benthic communities". Stream flow can vary with the frequency of rainfall, catchment slope, the physical properties of the land surface and the presence of lakes.

Rivers draining lakes and those fed from springs or similar groundwater sources, as in the Central North Island pumice country, have the least variable flow (Duncan & Jowett, 1990). Slightly more flow variations, is shown by rivers whose catchments are subject to regular but not constant precipitation, resulting in relatively constant baseflow. South Island West Coast rivers fell into this category. The greatest flow variation occurs in rivers draining areas with irregular precipitation, with correspondingly greater variation in runoff and baseflow. Rivers draining the drier eastern areas of both islands, and in particular the dry areas of South Canterbury, North Otago and Central Otago, have the greatest flow variability in New Zealand.

New Zealand lacks the very large catchments and rivers found in continental areas like Australia and North America. The Clutha River with a mean flow of 570 m^3/s is the largest river in New Zealand and the Waikato (327 m^3/s) the largest in the North Island. New Zealand has seven rivers (Table 1.1) with mean annual discharges greater than 200 m^3/s.

New Zealand's rivers tend to have very large flows (high specific discharges) for the size of their catchments. Parde (1966) noted that the very large flood flows from West Coast rivers are related to their steep mountainous catchments, to the lack of lake storage, and to the extraordinarily high precipitation, of up to 12,000 mm/per year.

The McKerchar & Pearson (1989) report on flood frequency in New Zealand provides a map-based method of estimating flood flows for ungauged sites, and also contains summaries of all the flood data used in preparing their report.

River	Catchment Area (km^2)	Mean Annual Discharge (m^3/s)
Clutha	20,582	570
Waiau	8,134	437
Buller	6,350	428
Waitaki	9,760	367
Grey	3,830	337
Waikato	11,395	327
Wanganui	6,643	224

Table 1.1 Flows of Larger Rivers in New Zealand Source: Faces of the River (Young & Foster, 1986) and Walter (1989).

Water Use in New Zealand

New Zealand is well endowed with fresh water, but competition between those who wish to use the water, and those concerned with preserving the rivers in their natural state, has increased markedly since the passage of the Water and Soil Conservation Act in 1967. This legislation, operative from 1967 to 1991, sought to promote multiple use of water resources. It has been replaced by the Resource Management Act of 1991, which has a rather different philosophy. The purpose of the Resource Management Act is "to promote the sustainable management of natural and physical resources".

New Zealand's water resources have been extensively developed for irrigation, hydro-electricity generation, and water supply. At the same time, a growing number of rivers, lakes, and wetlands are being preserved from development, through Water Conservation Orders or because of their location in national parks. The creation of new parks, such as Paparoa National Park on the West Coast, provides a measure of protection for the rivers within their boundaries.

Table 1.2 provides data on average and maximum daily water use for some towns and cities in New Zealand. The data supplied by the water supply agencies are largely for residential use but in some cases include some industrial use and leakage from the reticulation system. It is clear that water use varies a great deal, perhaps reflecting the widespread view that there is plenty of water available in New Zealand. Peak demand in residential areas normally occurs in summer with widespread use of sprinklers to water gardens and lawns. To date there have been few attempts to conserve the use of water in urban areas, apart from short term restrictions on water use during periods of drought.

Catchment	Maximum daily use	Average daily use
Manukau City	430	N. A.
Dunedin City Council	769	497
Taupo District Council		
Central - West Zone	1371	610
South Zone	1760	601
Acacia Bay	1418	623
Gisborne	350	180
Marlborough		
Renwick	2000	760
Wairau Valley	1550	925
Havelock	720	400
Rarangi	390	195
Picton	1340	780
Blenheim	1200	810
Nelson	280	190
Greymouth		
meter	1080	400
non-meter	1620	600
Invercargill (includes industrial)	625	360
Kapiti Coast		
Waikanae	1007	674
Paraparaumu/Raumati	720	480
Otaki	940	495
Paekakariki	706	212
Christchurch	2000	450
Thames Valley		
Coramandel Resort Towns	1220	860
Opotiki District Council	709	497

* All units are in litres per person/day
* Water use is largely residential but in some cases incudes industrial use and leakage.

Table 1.2 Water use in some New Zealand Towns

Irrigation, which uses 1.1km^3/yr, is the major consumptive use of water in New Zealand. This underlies increasing conflict between water conservation interests and farmers needing irrigation water to support pastoral agriculture, cropping and horticulture.

Organisations and Structures

The recognition of the need to control the costly and dangerous rampages of New Zealand rivers is shown by the early passage of the River Boards Act in 1884. The 1908 River Boards Act is still the basis of river board work (Acheson, 1968) and with the Land Drainage Act (1908) controlled much of the early river and drainage work in New Zealand. The Soil Conservation and Rivers Control Act and its amendments widened the scope of earlier work to cover soil conservation and catchment control schemes. It also established Catchment Boards and a central agency, the Soil Conservation and Rivers Control Council (SC & RCC).

The Water and Soil Conservation Act (1967) established Regional Water Boards, and added water allocation and pollution control to the earlier powers of the Catchment Boards. In 1989, these organisations were replaced by Regional Councils with wider powers including regional planning, pest and weed control and Civil Defence functions. The National Water & Soil Conservation Authority was abolished and much authority was devolved to the Regional Councils.

The enactment of the Resource Management Act (1991) requires Regional Councils and Unitary Authorities to produce a regional policy statement for each region of New Zealand. Regional policy statements identify the major issues for the region and set out policy statements. The Resource Management Act also allows for national policy statements and requires the Minister of Conservation to produce a New Zealand coastal policy statement. This will establish national priorities for the preservation of the natural character of the coastal environment and deal with issues such as subdivision, use, or development of the coastal environment. Regional policy statements may not be inconsistent with, or in conflict with any national policy statement. Similarly, regional plans and catchment management plans have to comply with the policies set out in regional policy statements.

Regional Councils may produce catchment or water management plans to manage water resources. These plans now have statutory backing and there is provision for integrated management of resources. At all levels of resource planning the Resource Management Act requires or allows for public input to the planning process.

The Value of Water

People tend to take water for granted, and it is not bought and sold like many other commodities. As a result, it is difficult to place a monetary value on New Zealand's water resources. Table 1.3 shows estimates of the value of some of these uses.

Activity	Value ($million)
Water supply (agriculture, industry, domestic)	450
Waste Disposal	450
Freshwater Fisheries	100
Recreation and Amenity Values	500
Hydro electric Power Generation and Thermal Plant Cooling	800
Gravel Resource replenishment	40
Total	2,340

Table 1.3 Estimated Economic Value of New Zealand's Water Resources. Source: Mosley, 1988.

This, of course, is in addition to the more intangible, but nonetheless real, values put on water by many different sectors of the community.

To Maori, water is the essential ingredient of life - a priceless treasure left by ancestors for the life -

sustaining use of their descendants. Taylor & Patrick (1987) gave an indication of what water means to the Maori and non-Maori inhabitants of New Zealand (Figure 1.2).

NON-MAORI	MAORI
Recreation	
Wilderness	
Meditation	
Cleanliness	
Beauty	
Power	
Wealth	
Nature	
Spirituality	
Food	
	Tradition
	Mana
	Alive

Figure 1.2 The Value of Water

The practice of discharging effluent into rivers, estuaries and the sea is not acceptable to Maori communities, particularly where the waters are used for traditional food-gathering, eg. eel fishing, shell-fish gathering. Indeed, there has been growing opposition throughout the community to waste disposal into fresh and coastal waters, and a readiness to adopt more expensive methods of treatment. Widespread adoption of such attitudes towards water might well provide a major impetus for better use and management of water resources in the 1990s.

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2
Flow Regimes of New Zealand Rivers

M J Duncan

In New Zealand you can see a wide range of types of river. There are boulder-filled mountain torrents, issuing from glaciers in mountains only a few kilometres from the coast; wide, braided, gravel-bedded channels; meandering silt watercourses; tree-lined urban waterways. Some which rise in the high mountains may change dramatically along their course, before discharging via a lagoon or estuary to the sea. Rivers arising in the foothills or from lowland springs tend to be more uniform.

But every river has its own unique character. What makes each one different? The answer lies in the combination of physical and climatic features that influence what we call the "flow regime" or "hydrologic regime" of a river.

What Is Flow Regime?

The flow regime (or hydrologic regime) of a river is the unique way that its flow changes from day to day, season to season and from one year to another. Regime defines the character of a river; how liable it is to flood or to experience long periods of low flow; what it looks like; whether it is potentially useful.

Differences in flow regime are best illustrated by looking at graphs of flow from different rivers. A hydrograph is a graph of the change in either a river's water level (often called stage) or its flow (discharge) over time. Two main components of river flow can be identified from a hydrograph: baseflow and flood flows (often termed quickflow) (Figure 2.1). The baseflow of a river is derived from seepage of ground water into the channel or from lake outflows; it may be large or small, but it tends to change slowly. Flood flows occur on top of the baseflow. They are produced from precipitation directly into the channel, from overland flow down surfaces sloping into the channel, from water that infiltrates into the soil and moves quickly to the stream channel (interflow) and from runoff from wet areas near stream channels. (See Chapter 15 for more detail).

Hydrographs of floods commonly show the rise of flood waters (termed the "rising limb") and their recession ("falling limb") (Figure 2.1). The slopes of the rising and falling limbs tell us about the nature of the rainfall that caused the flood and the catchment itself (see Chapter 6). For example, during "flash" floods caused by intense rain falling onto an already saturated catchment streams rise rapidly because the water runs directly into the stream network rather than soaking into the ground.

In hydrograph analysis, baseflow and floodflow are separated by drawing a line from the start of the rising limb of a flood to a point on the falling limb. Because baseflow is usually higher after a flood, the line has an upward slope which is usually chosen from experience; a figure of 0.004 l s^{-1} km^{-2} (Hewlett and Hibbert 1967) was used to separate the flows in Figure 2.2.

The particular combination of baseflows and flood flows for a river form its flow regime. Rivers may have a stable regime with limited variation in

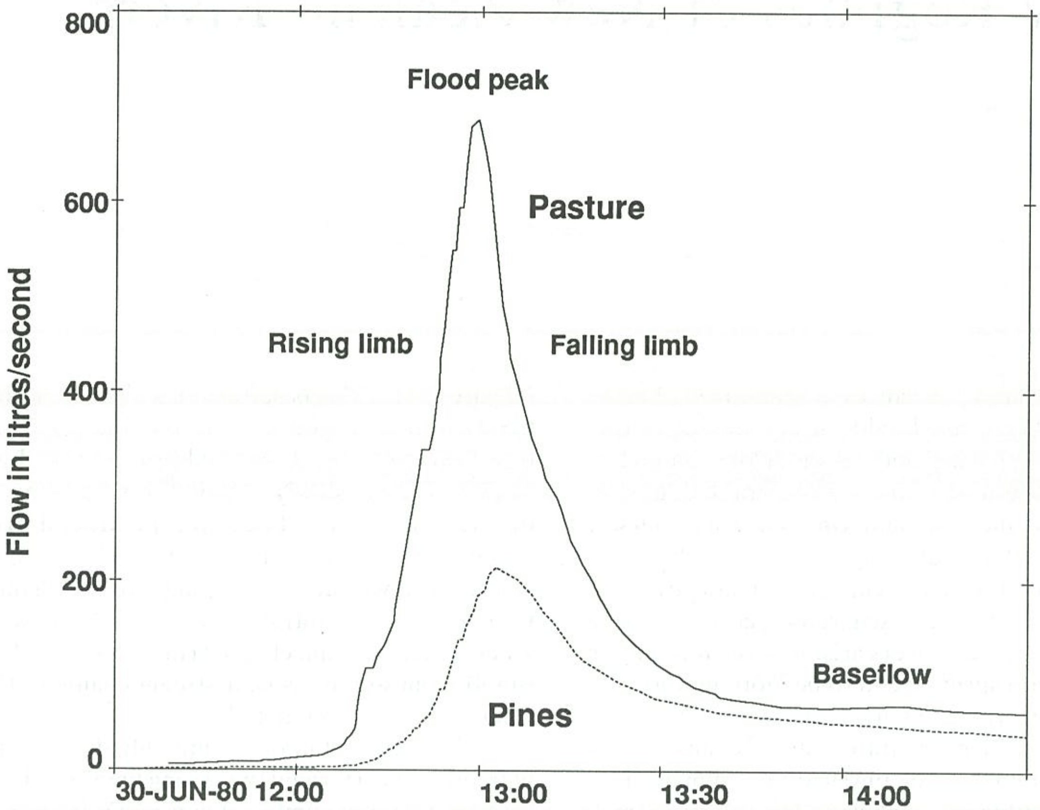

Figure 2.1 Hydrograph showing the rising limb, and falling limb of the flood and the baseflow. The difference in regime caused by land-use change is shown by hydrographs from a pasture catchment (—) and a pine catchment (...) of similar size (approximately 7 hectares) for the same rain storm.

flow, or a regime with very variable flows. The Buller River at the outlet of Lake Rotoiti (Figure 2.2A) shows small, regular, slowly rising and falling floods on top of a large sustained baseflow. The opposite extreme is illustrated by the Whareama River (Figure 2.2D) which shows a clear seasonal pattern of virtually zero flow in summer but a sustained baseflow in winter, with frequent, short, flashy floods.

Floods may happen regularly, e.g., virtually weekly on the South Island's West Coast, or only occasionally. In some rivers, floods are seasonal: often in winter and spring in east coast streams.

Why Is The Flow Regime Important?

The River as a Habitat

The flow regime of a river, in combination with other factors such as temperature and water quality, influences the plants and animals that can live in it. As an example, consider the conditions favorable for brown trout. They like cool, clear, bouldery, rivers with stable flow regimes with few floods and high baseflows (Chapter 14 - Jowett, 1990). There are several reasons for this. When the river bed is nearly always covered in water the

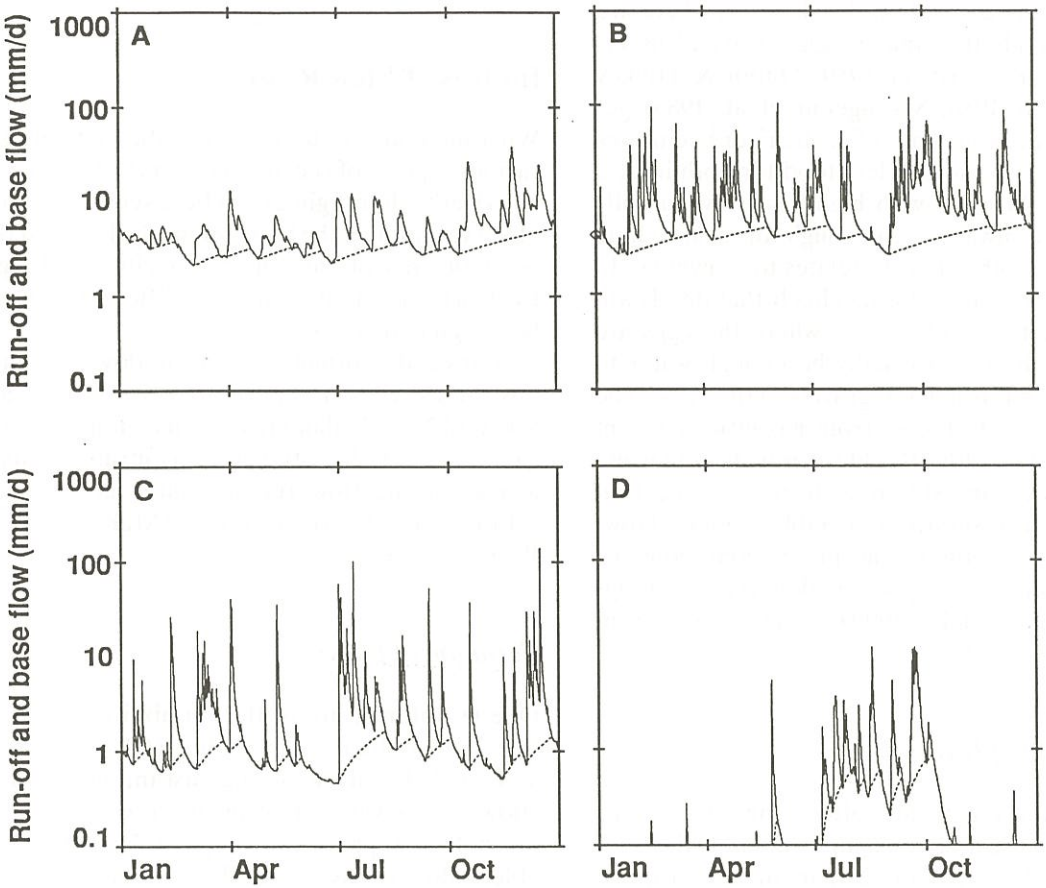

Figure 2.2 Examples of annual runoff and baseflow hydrographs of rivers with low- and high- variability flow regimes. Note the logarithmic scale. The dotted line separates baseflows from flood flows using the method of Hewlett and Hibbert (1967) (see text.)

Graph	River	Year	Catchment Area(km^2)	CV
A.	Buller at Lake Rotoiti (Nelson)	1984	195	0.76
B.	Ahaura River (Westland)	1981	790	1.06
C.	Wairoa River (Nelson)	1984	464	2.40
D.	Whareama River (Wairarapa)	1984	398	3.37

(After Jowett and Duncan 1990).

food chain can maintain full production. Algal slimes can grow on the gravels and boulders on the river bed, and aquatic insects, the main food of trout, in turn can feed on the slimes. If there are frequent floods, the slimes (Biggs 1990) and insects (Jowett & Richardson 1989, Quinn & Hickey 1990b, Sagar 1986, Scrimgeour et al. 1988) get washed away or ground off as the river bed moves in the flood, so there is less food for both insects and trout. Streams with high base flows usually water that is always deep enough for trout to hide and rest. Another reason relates to spawning. In a stable flow regime, it is less likely that floods will wash away the redds (areas where the eggs are laid), and there will usually be enough water to carry oxygen through the gravels to the eggs. The clear water preferred by trout has little sediment to clog up the redds. Boulders provide both white water to hide, and still areas to rest. Thus, trout are adapted to survive in a stable regime. However, other creatures may prefer conditions associated with other types of flow regime (Biggs 1990, Quinn & Hickey 1990a,b, Sagar 1986, Scrimgeour et al. 1988).

Human Use of Rivers

A river's flow regime also affects the way in which people can use it. For example, monthly flows for the Rakaia River (Figure 2.3) are highest in spring and summer. This is also the time of highest demand for irrigation water. Therefore, water can be taken directly out of the river and there is no need for costly storage reservoirs. On the other hand, demand for hydro-electric power peaks in mid-winter, and expensive dams have been built to augment the storage capacity of lakes such as Tekapo, Pukaki and Hawea. They have river inflow patterns similar to those of the Rakaia and Ahuriri Rivers (Figure 2.3).

Sometimes a river's flow regime is very suitable for one use but poor for another. For example, Nelson rivers have a monthly flow regime similar to that of the Hakataramea River in South Canterbury, (Figure 2.3) where summer flows are low, with slow clear water. These conditions are ideal for swimming, and suit holiday makers visiting Nelson. On the other hand this same flow regime restricts the amount of water available to irrigate Nelson's important horticultural crops.

Indices of Flow Regime

What methods are used to describe and compare various aspects of (i.e. to "index") the flow regime of a river? Flow regimes can be discussed in terms of the variation of the flows, monthly flow, and flow per unit area of catchment (specific discharge). Each of these tells us something different about the flow regime of a river.

Maps and information on mean flows, sediment discharges, river temperatures, low flows and floods of New Zealand rivers can be found in Duncan (1987). Other studies concentrate on single aspect of the flow regime, such as low flows (Hutchinson 1990) or floods (McKerchar and Pearson 1989)

Variation of Flows

One useful measure of the variability of a river's flow is its coefficient of variation (CV). This is the standard deviation of the instantaneous flows (flows at a given instant in time), divided by the mean flow. It provides a simple index of how variable a flow is. An example of its application is found in Jowett & Duncan (1990). They examined flow regimes of New Zealand rivers using records of instantaneous flow, usually recorded every 15 minutes, for 130 river sites. They classified the sites according to indices describing flow variation, including the magnitude and variation of high and low flows. They also identified the catchment characteristics which contributed to flow variability.

Each of the 130 sites was classified into one of six groups based on flow variability. In general, as flow variability increased, so too did the variation in base flow, the magnitude of flow fluctuation above baseflow, and the flow recession rates. In fact all the flow variability indices correlated well with each other. However, the coefficient of variation of flow probably best described the flow regime.

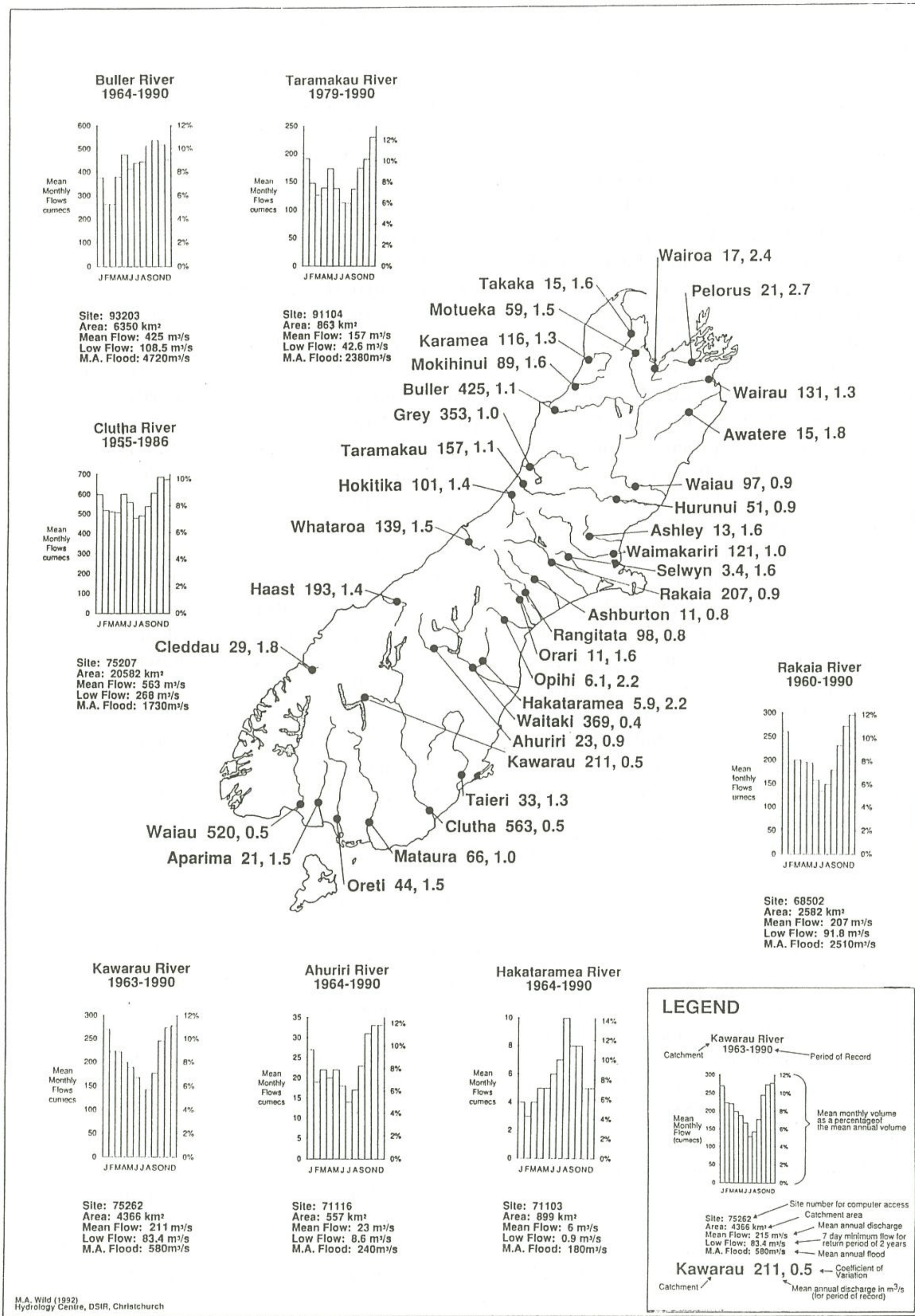

Figure 2.3 Map of the mean flow and coefficient of variation of major rivers in the South Island, based on data held in National Institute of Water and Atmospheric Research or Regional Council archives.

Low variability of flow indicates a stable flow regime. Rivers with little variation in flow (CV < 0.85) tend to be controlled by large lakes or are mainly spring-fed, such as the Buller River at Lake Rotoiti (Figure 2.2A). Their flow is mainly baseflow, and there are very few, small floods. Such rivers are rich in nutrients and they normally support a large amount of stream life. Rivers with slightly more variable flow (CV = 0.85 - 1.25) tend to drain high rainfall areas; they have a high baseflow, but also have frequent, large floods which disturb the river bed. These conditions do not allow stream plant and animal communities to develop fully (Biggs et al. 1990). West Coast rivers such as the Ahaura (Figure 2.2B) fall into this category. Variations in flow are greatest (CV >3) for rivers subject to irregular precipitation. These rivers have long periods of low flow, low base flows, and large infrequent floods of short duration, e.g., the Whareama River (Figure 2.2D). Rivers with CV's in the range 1.25-3 have intermediate characteristics, e.g., Wairoa River Figure 2.2C).

Figures 2.3 and 2.4 show the mean flows and coefficients of variation of flow of sixty of the country's larger or more economically important rivers. Some large South Island West Coast rivers and large east coast rivers such as the Clarence River in Marlborough are not included because their flows have not been reliably measured. The mean flows shown are the natural river flows, i.e., they are the flows that would be expected if there were no man-made diversion of flow from one catchment to another.

Monthly Flow Histograms

Month-to-month variations in river flow Figures 2.3 and 2.4) primarily reflect the seasonal distribution of rainfall in New Zealand. The winter rainfall peaks in the north are reflected in the flows of the Awanui, Motu, Wanganui and Manawatu rivers, while the more even distribution of rainfall in central New Zealand is illustrated by the flow of the Buller river.

The monthly flows of the Rakaia, Kawarau and Ahuriri rivers are typical of alpine snow-fed rivers, where winter precipitation is held in the snow pack and released in the spring and summer thaw. However, the high spring and summer flows are also a response to rainfall from the northwest winds which prevail then. The Taramakau River flow pattern, which is typical of the many short, steep and large rivers draining the Southern Alps to the west, shows the same traits but is less influenced by snow melt.

In the Volcanic Plateau of the central North Island, rainfall percolates through the fractured pumice into the groundwater system and is released evenly by spring-fed streams, as in the Tarawera River (Figure 2.4). Many of the rivers with headwaters in the central North Island show some influences of their pumice cover. A comparison of the Waipaoa River, which drains tertiary sedimentary rocks, with the Rangitikei or Wanganui River shows how much the flows are moderated in the latter rivers.

Lake storage and the even release of water for hydro-electric power generation are responsible for the uniform monthly flow of the Clutha at Clyde (Figure 2.3). Its monthly flows vary much less than those of its tributary the Kawarau, even though Kawarau monthly flows are somewhat moderated by uncontrolled storage in Lake Wakatipu. The Buller River flows also shows the smoothing influence of lakes Rotoiti and Rotorua on its monthly flow fluctuations.

The Hakataramea River is typical of foothills-fed east coast rivers, with high flows in late winter and spring and low flows in summer and autumn. These reflect the generally low east coast rainfall, and dry summers when the soil dries out. Only when soil moisture is fully replenished by a combination of winter rainfall and low rates of evapotranspiration, is there sufficient rainfall to increase flows substantially.

Specific Discharge

Specific discharge (also known as specific yield) is the flow per unit of catchment area, usually expressed in litres per second per square kilometre ($l\,s^{-1}\,km^{-2}$). It allows the flows from different catchments to be directly compared. It can also be converted to depth of runoff in millimetres, and is

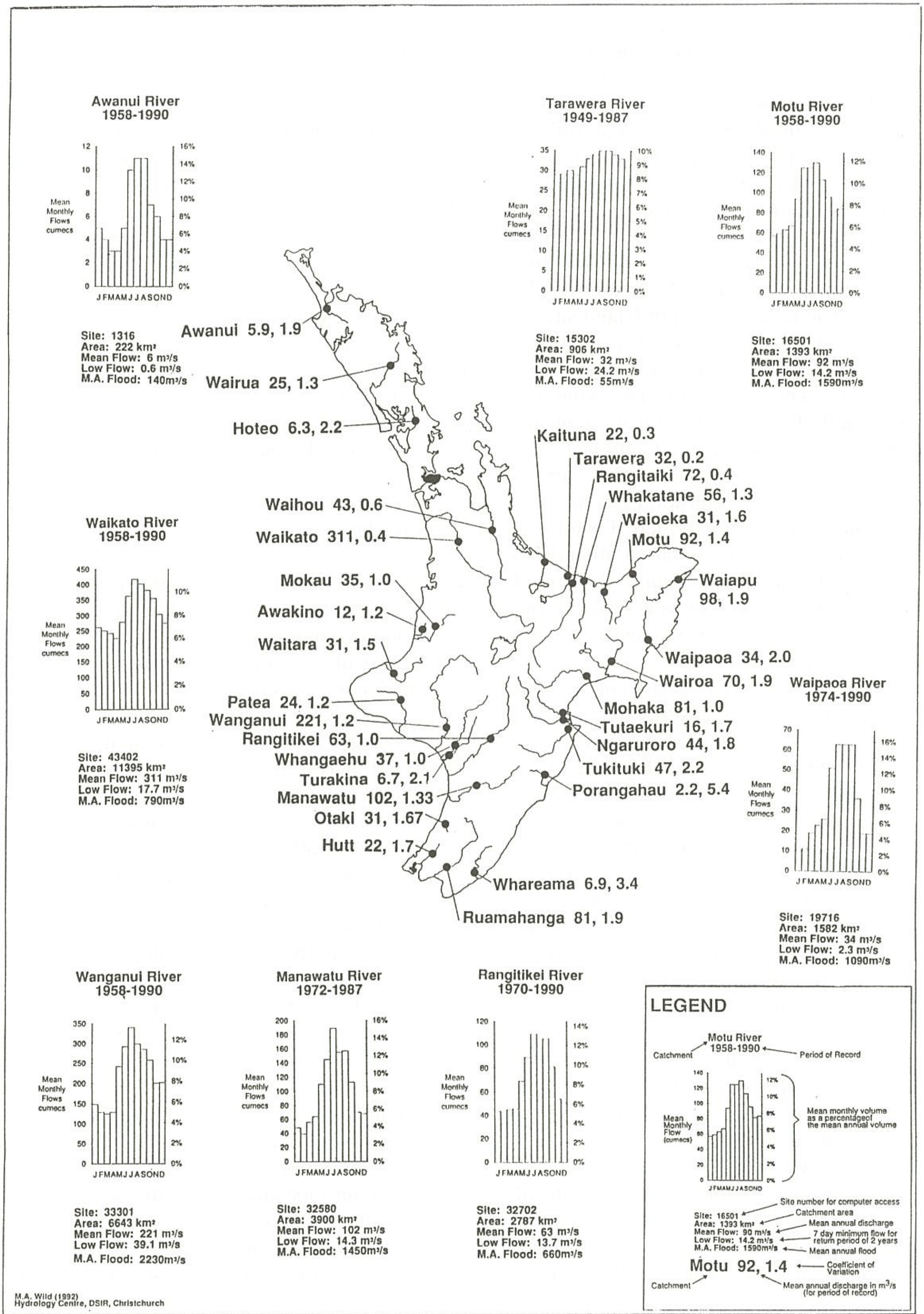

Figure 2.4 Map of the mean flow and coefficient of variation of major rivers in the North Island, based on data held in National Institute of Water and Atmospheric Research or Regional Council archives.

therefore easily compared with rainfall. Further insights into a river's regime may be made by examining its specific discharge for various parts of its flow, such as low flow, mean flow or flood flow.

The mean specific discharge strongly reflects catchment rainfall. The range for the North Island rivers shown in Figure 2.4 is 8 to 101 $ls^{-1}\ km^{-2}$ (290 mm and 3190 mm) for the Porangahau and Otaki Rivers respectively. However, most of the catchments yield about 34 $l\ s^{-1}\ km^{-2}$ (1070 mm) reflecting the relatively even distribution of rainfall over the North Island. The range for South Island sites is much wider, with the Whataroa River yielding a high 310 $l\ s^{-1}\ km^{-2}$ (9840 mm) and the Hakataramea River only 6.7 $l\ s^{-1}\ km^{-2}$ (210 mm). The Whataroa catchment runoff of over 9840 mm is by no means uncommon, as the Hokitika at Colliers Creek (catchment area 352 km^2) yields 8700 mm of runoff. To this, evapotranspiration estimates of 600-700 mm per year must be added (Finkelstein 1961) to indicate an annual rainfall of about 9500 mm over the whole catchment. Figures 2.3 and 2.4 show data for the major rivers of New Zealand, and flow rates for 96 smaller river sites can be found in Close and Davies-Colley (1990).

Specific (i.e. per unit area) mean annual flood flows reflect storm rainfall intensities that normally increase with annual rainfall. However, the highest rates for the North Island (Figure 2.4) are for rivers towards the north, which are subject to storms originating from tropical cyclones. These are rivers such as the Awanui, Motu and Waipaoa, which have specific mean annual floods of 630, 1140 and 690 $l\ s^{-1}\ km^{-2}$ respectively. Many of the other North Island rivers have specific flood flows of about 300 $l\ s^{-1}\ km^{-2}$. The Waikato and Tarawera Rivers have very low specific mean flood rates of 60-70 $l\ s^{-1}\ km^{-2}$, because lake and ground water storage in the pumice of the central volcanic plateau have strongly modified the flood regime. The annual maximum floods and related statistics for 343 rivers nationwide can be found in McKerchar and Pearson (1989).

Low flows are determined by recent rainfalls, catchment groundwater storage and its rate of seepage (a function of the underlying rocks and lakes), and catchment area. The lowest average flow over 7 consecutive days that could be expected to occur on average every 2 years is called the specific 2-year return period 7-day low flow. Such flows vary from about 50 $l\ s^{-1}\ km^{-2}$ in the Taramakau River to as little as 1 $l\ s^{-1}\ km^{-2}$ in the Hakataramea River, and are primarily a function of annual rainfall. Catchments with small low flows also tend to have long periods with low flows. Hutchinson (1990) lists low flow magnitude and frequency from 428 sites nationwide.

Shaping Flow Regimes

Flow regimes principally reflect climate - precipitation and evapotranspiration, geology, vegetation cover, and human activity such as flow diversion for hydroelectricity generation.

Climatic Influences

The major climatic factors influencing flow regimes are how often, and how hard it rains. Examination of the rainfall pattern helps to explain why a particular river has a particular regime.

New Zealand's rainfall pattern results from its long narrow shape, steep topography and isolated island position. The country's mountain backbone lies directly across the path of the eastward moving anticyclones and low pressure troughs which are characteristic of the "Roaring Forties". The passage of these weather systems results in a high and regular rainfall over much of the country, although some places get much more rain, more often, than others (Heine 1985). Mean annual rainfall varies from as little as 300 mm in a small area of Central Otago to over 10,000 mm in a long narrow strip to the west of the crest of the Southern Alps (Griffiths & McSaveney 1983). However, over most of the country it is between 600 and 1500 mm. Some areas with average rainfall under 600 mm are found in the South Island to the east of the main ranges. North Island mountains are lower, and annual rainfall is more uniform. Much of the island receives about 1500 mm and the dry areas (central and southern Hawke Bay, Wairarapa and Manawatu) about 700 mm (See Chapter 4).

Reflecting these differences in rainfall, then, rivers draining westwards from the Southern Alps have annual runoffs of the order of 5000 mm, whereas those draining the Wairarapa have annual runoffs of the order of 300 mm.

The greatest seasonal contrast in rainfall occurs in the north, where winter rain is almost double that of summer. The resultant effect on stream flows is evident from the patterns of monthly flow of the Awanui, Motu and Waipaoa River (Figure 2.4). This predominance of winter rainfall diminishes southwards (Tomlinson 1976), although it is still discernible over the northern part of the South Island and its effect can be seen in the flow of the Buller River (Figures 2.2A and 2.3). Further south, winter is the season with lowest rainfall, and inland areas receive most rainfall in summer, from convective showers. The effect of low winter rainfall can be seen in the Taramakau River (Figure 2.3), but the higher summer rainfall of the coastal catchment is more commonly due to northwest rainfall than convective showers. The highest variations in seasonal rainfall from year to year are in areas to the east of the mountain ranges. Here very dry conditions may develop in late summer and autumn, particularly in Hawke Bay, Canterbury and North Otago. The Hakataramea River monthly flows (Figure 2.3) and the Whareama River hydrograph (Figure 2.2D) illustrate the effect of these high seasonal variations in rainfall.

Usually it rains hardest where it rains the most (Tomlinson 1980; Whitehouse 1985). The highest 24-hour rainfall on record is 758 mm, which fell at Prices Flat in the Hokitika catchment, in the high rainfall zone of the western Southern Alps (Henderson 1991). A storage raingauge at Alex Knob on the south bank of the Waiho River, Fox Glacier, recorded 1800 mm in 3 days in March 1982. If rainfall at Alex Knob has a similar intensity pattern to that at neighbouring recording raingauges, and we think it does, about 1350 mm would have fallen in 24 hours (Thompson and McKerchar 1992). Such high and intense rainfall produces frequent flashy floods imposed upon a sustained baseflow, as is evident in the hydrograph of the Ahaura River (Figure 2.2B).

The Gisborne and Auckland regions, which have considerably lower annual rainfalls than the Alps, can also receive heavy daily falls of as much as 140 mm. In contrast, Otago and Southland rarely receive daily falls greater than 110 mm and 80 mm respectively (Thompson 1987).

Geological Influences

Some types of rock allow water to pass through them much more easily than others. That is, their transmissivity (defined as the rate at which water moves through the ground) is higher. Certain combinations of rock provide ideal conditions for the storage of water as groundwater. This implies that the type of rock, or the lithology, in a catchment affects the rate at which rainfall flows over and through the catchment to the river. For example, rocks such as Tertiary mudstones, shales and siltstones which have low transmissivity and little storage tend to produce flow regimes which have flashy floods, which recede quickly and low base flows. Rocks of this type occur in the Whareama River catchment, in the Wairarapa (Figure 2.2D). Catchments with high infiltration (defined as the movement of surface water into the soil and rocks), transmissivity and water storage tend to have small floods with slowly receding flow, and large and persistent baseflows. Examples are the Maryburn in the McKenzie Basin which has deep permeable gravels at the surface, or the Rangitaiki River (Figure 2.4) which drains an area with a deep pumice cover.

In his study of summer low flows in Northland, Waugh (1970) found that fissured basaltic lava absorbed rainfall and released it slowly, thus sustaining low flows. Areas with other rock types such as Cretaceous shale and sandstone were less absorbent, and their streams had lower low flows. A study of water resources of the Nelson area (Scarf, 1972) showed that the rivers draining from the marbles of the Mt Arthur Range had substantial low flows, some issuing from caves (e.g., Riwaka River) and springs, e.g., Pupu Springs, Takaka). This was in contrast to the very low flows of streams draining areas covered by the impervious Moutere outwash gravels, where streams commonly dry up in summer. Although rainfall distribution plays a part, catchment geology has a major influence on

Nelson flow regimes.

Lake storage has an effect on flow regimes which is similar to that of rocks with high storage characteristics. For example the Buller River at the outlet of Lake Rotoiti (Figure 2.2A) shows flow peaks which are much more subdued than those of the Ahaura River, because of the damping effect of the lake.

Human Influences

In many New Zealand rivers the natural flow regime has been altered, particularly by hydro-electric power projects or changes in land use. Hydro-electric development has substantially affected the Waiau (Southland), Wanganui, Clutha and Waitaki Rivers. The mean flow of the Waiau River has been reduced from 520 to 175 $m^3 s^{-1}$ by the 345 $m^3 s^{-1}$ of flow which has been diverted to Doubtful Sound via the Manapouri Power Station. (The reported natural mean flow of the Waiau of 520 $m^3 s^{-1}$ (Figure 2.3) was obtained by combining Waiau and Manapouri flows). The flow of the Waiau River at Tuatapere before and after diversion (Figure 2.5) show the flow regime has been affected by a reduction in the full range of flows.

Some of the headwater streams of the Wanganui River have been diverted into the top of the Waikato system - much of their low and median flows are now redirected, leaving only small residual flows and flood flows. However, the normal regime of the Wanganui is partially restored as undiverted tributaries add to its flow. Hydrographs

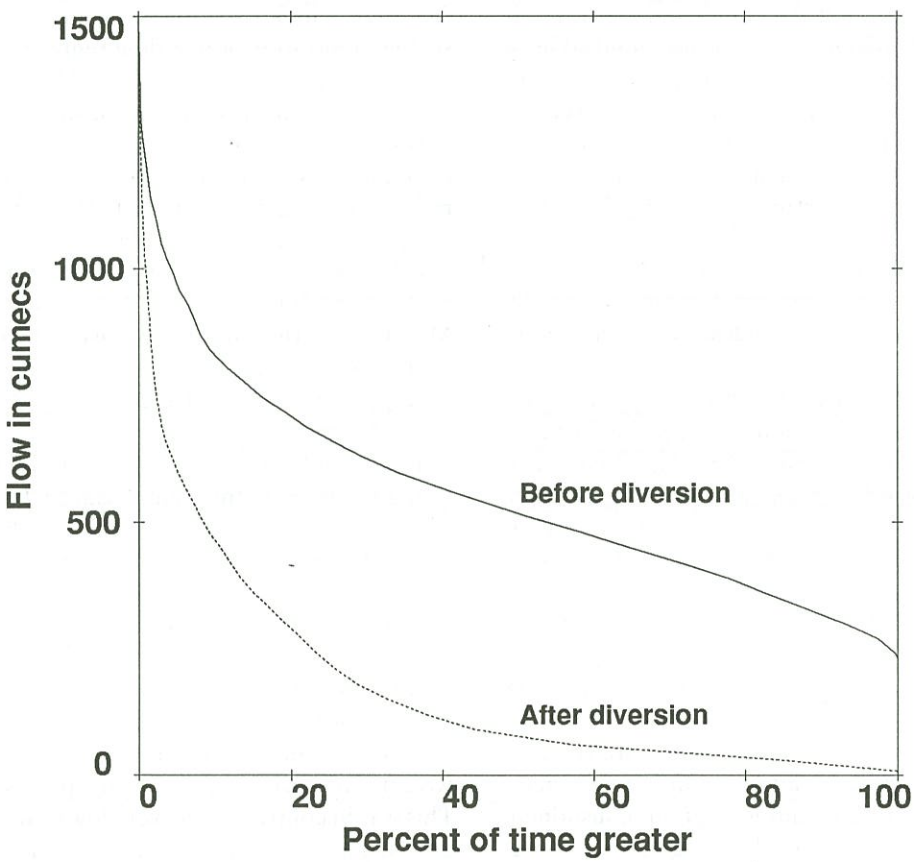

Figure 2.5 Flow duration curves of the flow in the Waiau River at Tuatapere before and after diversion via Lake Manapouri to Deep Cove.

of the remaining flow and simulation of the natural flow of the Wanganui River at Te Maire (Figure 2.6), illustrate that the low flow part of the flow regime is most affected by the diversion. The loss of 18 $m^3 s^{-1}$ from the Wanganui River system is the Waikato River's gain. The Waikato also gains 14 $m^3 s^{-1}$ from the Rangitikei River via the Moawhango Tunnel. It is further modified by controlled outflows from Lake Taupo and eight hydro-electric power stations further downstream. The net effect has been to reduce flood flows and increase low flows in the Waikato River.

At the Roxburgh hydroelectric power station the release of extra water from the lake to meet peak electricity demands causes a daily flood wave on the Clutha River (Figure 2.7). It has been suggested (Otago Catchment Board 1986) that this, combined with the tidal and wave pattern at the coast, has resulted in periodic shifts of the river mouth, leading to regular flooding in the Lower Clutha delta. Future dams between Roxburgh and Balclutha may yet moderate these daily flow fluctuations. The monthly Clutha flows (Figure 2.3) mask the daily fluctuations. The monthly regime is even because of the moderating effects of the large lakes Wakatipu and Wanaka and the manipulation of water storage in Lake Hawea.

Hydro-electric storage dams and diversion canals in the Waitaki Catchment have made dramatic changes to the flow regimes of its large rivers. The Ohau River previously had a mean flow of 80 $m^3 s^{-1}$, but now has either no flow or occasional flood flows. However, recently agreement has been reached on releasing a residual flow of 10 $m^3 s^{-1}$ in exchange for being able to operate Lake Ohau over a larger range of lake levels. The Pukaki River now has no flow. Apart from occasional flood flows there is only a small release of

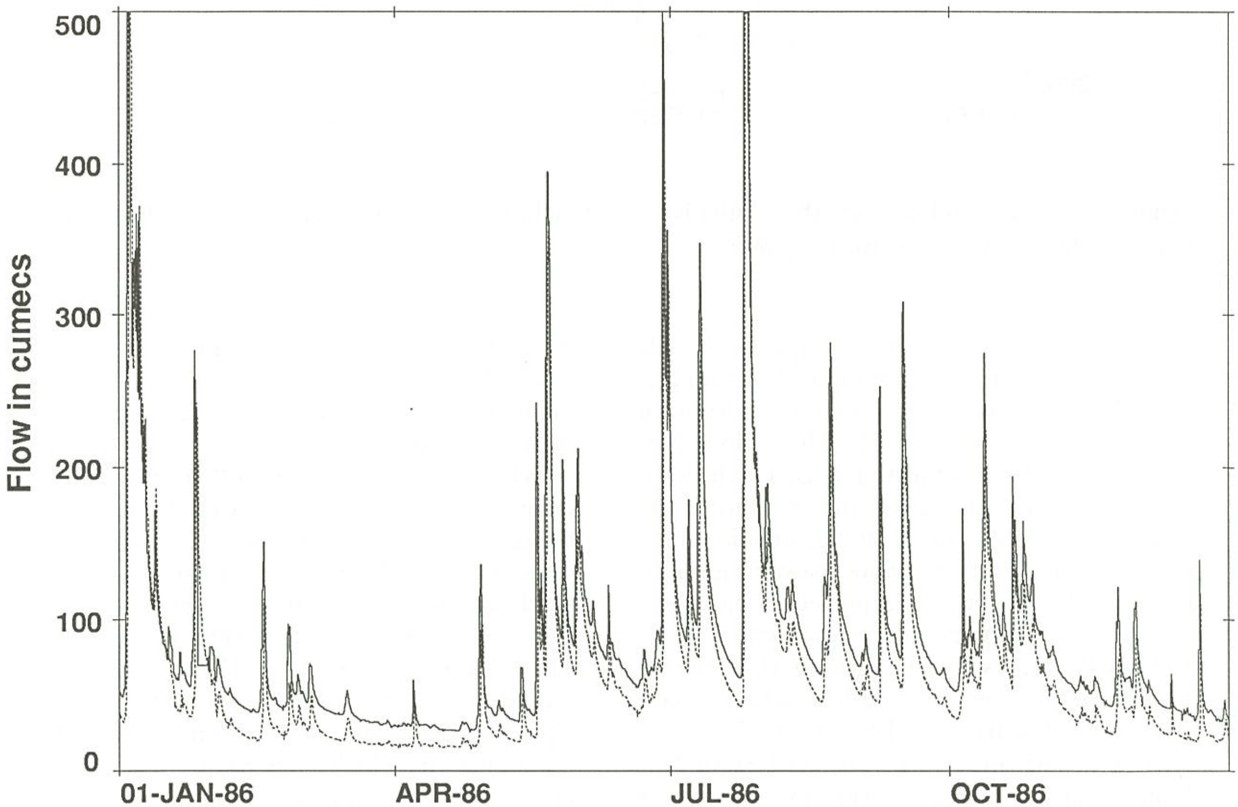

Figure 2.6 Hydrograph of the residual flow in the Wanganui River at Te Maire after diversion of its headwaters (dotted line) into the Waikato River, and a simulated hydrograph of natural flows (solid line).

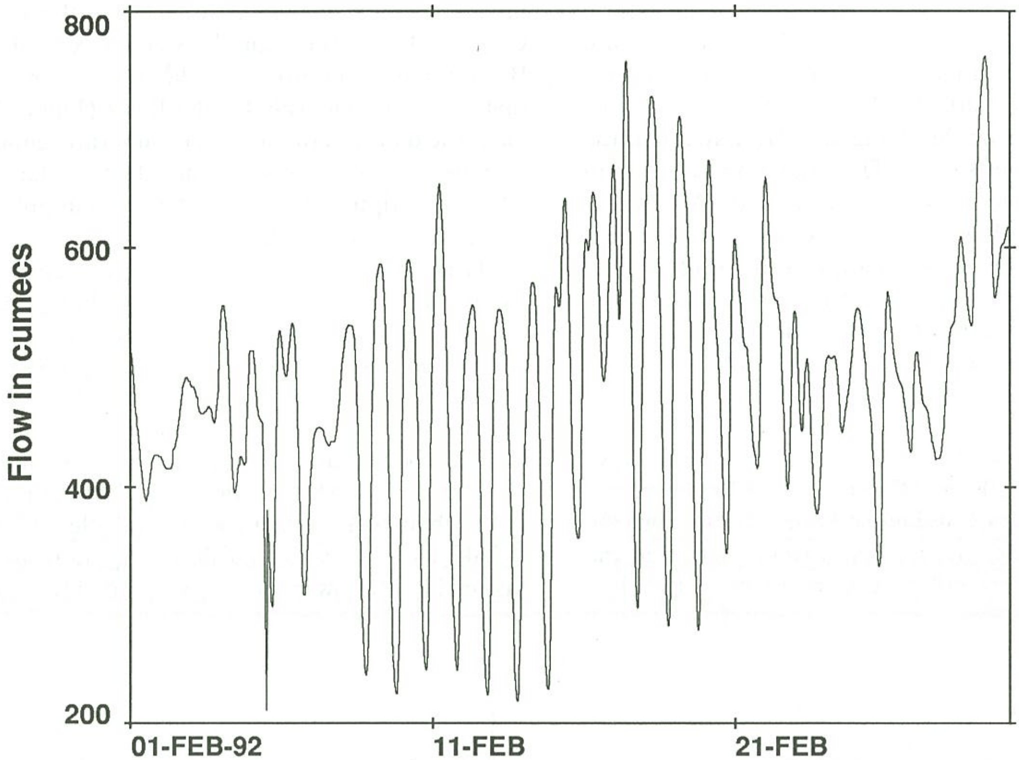

Figure 2.7 Daily flood waves in the Clutha River at Balclutha caused by Roxburgh hydroelectric power station responding to demands for power.

turbid glacial water into the Tekapo River. The river further downstream is now much clearer than before and conditions for trout have been enhanced (Teirney et al. 1982). Flood flows have been reduced and low flows increased in the lower Waitaki River. However, the Roxburgh and Waitaki dams have prevented Chinook salmon from returning to their previous spawning grounds, and the salmon runs are reported to have been substantially reduced (Teirney 1980). More biological effects of changes in flow regime downstream of New Zealand hydroelectric dams are described by Irvine and Jowett (1987).

Both Maori and European settlers in New Zealand have influenced river flow regimes by making large-scale modifications to the vegetation. The moa hunters effectively converted large areas of bush to tussock country by burning. Europeans in turn have converted tussock and bush country to pastoral farms, and bush and scrub country to pine plantations. Chapter 15 details the effects of some of these changes.

When land is cleared, runoff from the land increases markedly, thus increasing floods and low flows. When pine plantations replace pasture, flood peaks may decrease by 80% and annual yields and low flows can halve (Figure 2.1); the opposite happens when pasture replaces pines. These changes occur primarily because of differences in interception of rainfall by vegetation types. Interception is the rain which falls on vegetation and which evaporates before reaching the ground. Thus it is not available for transpiration by the plants or for runoff. Interception in tall vegetation like trees or scrub can account for 20-30% of rainfall.

Estimating Flow Regimes of Ungauged Catchments

Streamflows have been measured for only a limited number of rivers and streams in New Zealand. It is often, however, necessary to estimate the magnitude of floods and low flows for catchments which are not gauged. To do this, hydrologists have tried to define regions in which basins are sufficiently alike to apply the measured relationships between rainfall and runoff from gauged basins to ungauged basins within that region. The high variability of geology, topography, and especially rainfall in New Zealand have made this a difficult task.

Toebes and Palmer (1969) divided New Zealand into 90 regions based on geology and climate, and proposed that representative basins monitoring rainfall and runoff, be established in each region. Fifty-three regional basins (Duncan 1987) were instrumented and, with those rivers instrumented for flood warning and power or irrigation development, served as the basis for flow estimates.

Beable and McKerchar (1982) proposed regions for the estimation of flood size and frequencies based on regional equations. They defined 7 and 6 regions, respectively, in the North Island and 6 and 3 regions in the South Island. While this was a useful exercise, difficulties arose at regional boundaries where widely varying flood estimates could be made depending on which regional equation was adopted.

In the North Island, where regional geology and soils vary more than in the South Island, cluster analysis suggests that useful regions cannot be easily identified (Mosley, 1981).

A more recent study, using a larger data set and longer records, demonstrated that contour maps of mean annual floods and 100-year floods could be drawn for the whole country (McKerchar and Pearson, 1989). This shows that flow regimes vary smoothly across New Zealand, rather than abruptly changing at well-defined regional boundaries.

Equations for estimating the low flow of ungauged catchments based on 11 regions nationwide were proposed by Hutchinson (1990). Many regional equations were quite similar, with the differences justified by providing more precise estimates. Paradoxically the Southern Alps region and North Island Central volcanic plateau, regions of quite different geology and rainfall regime, had similar equations for the estimation of low flows. The regular Southern Alps rainfall and the porous volcanic plateau bedrock may both have the effect of sustaining low flows.

Flow variability was the basis for the classification of 130 river sites by Jowett & Duncan (1990) discussed previously. They did not attempt to identify hydrologic regions but they did map their six groups. Rivers with lowest flow variability were associated with the large South Island montane lakes, and the volcanic plateau of the North Island. The next group was also in the central portions of the North and South Islands. The group with the greatest flow variability is on the east coast of both islands, and an intermediate group included rivers around Mt Taranaki, the Tararua Ranges and in the Nelson region.

Rock type (soft and hard sediments, igneous, volcanic ash), flow variability and water quality (mainly conductivity), because of their links with biological communities, were the basis of regional groups of rivers to form riverine "ecoregions" in a study by Biggs et al. (1990). Five principal riverine ecoregions were distinguished. Particularly distinctive were the hydrological, geological and water quality conditions of the central North Island volcanic plateau and the eastern, Hawke Bay-Poverty Bay region of the North Island. Other regions were the Tararua Ranges, the remainder of the North Island comprising Taranaki, Waikato and Northland, and the South Island. These regions could have considerable benefits for establishing river management goals especially where unmanageable factors such as catchment geology may cause naturally poor water quality compared with other regions.

In summarising regional hydrological regimes Mosley (1981) stated that in the South Island, climatic regime, as modified by topography, appears to be the major influence. Much of the South Island is underlain by relatively impermeable rocks, and has steep topography. They are less important as sources of variation in flow regime than climate, which is spatially highly variable.

The North Island is more complex, with variations in flow regime influenced by climate (e.g., the Northland sites), lithology and soils (e.g., pumice area sites), and topography (e.g., sites draining the Tararua Range, and Mt Taranaki).

Summary

The flow regimes of New Zealand rivers are determined primarily by the country's abundant rainfall, steep topography and long narrow shape, which give rise in the main to short, swift gravel or cobble-bedded streams. Local weather conditions, geology, lakes and hydro-electricity schemes further modify flow regimes.

Coefficients of variation (CV) of flow together with mean flow, are a good measure of flow regime. Values of CV vary from about 0.55 for spring-fed streams to over 3 for some east coast streams. Specific mean flows, reflecting annual rainfall, vary from about 280 $l\ s^{-1}\ km^{-2}$ for South Island west coast rivers to less than 7 $l\ s^{-1}\ km^{-2}$ for small east coast rivers. Specific mean annual flood flows, which are influenced by rainfall intensity, catchment storage (lakes, lithology) and catchment area, vary from over 5000 $l\ s^{-1}\ km^{-2}$ to less than 60 $l\ s^{-1} km^{-2}$. Seven-day minimum flows with return periods of 2 years, determined primarily by rainfall regime and catchment storage, range from over 50 $l\ s^{-1}\ km^{-2}$ to less than 1 $l\ s^{-1}\ km^{-2}$.

Acknowledgements

I would like to thank Cathy Holmes, Eileen McSaveney and Paul Mosley for constructive reviews and Kathy Walter for preparation of figures. Figures 2.3 and 2.4 are adaptations of in idea of John Waugh's, first drafted by Doug McMillian and compiled here by Michelle Wild using data revised by Kathy Walter.

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3
Data, Information and Engineering Applications

M Paul Mosley, Ian Jowett, and Alaric I Tomlinson

Introduction

Data are a fundamental starting point for both the scientific study of hydrological processes and engineering applications to the planning, design, and operation of water resources projects. These projects include hydroelectric power, irrigation, industrial and municipal water supply, environmental protection, river control, rail and road transport, and land resource planning. Data requirements for each study or project may differ in detail, but there are many basic requirements and procedures in common.

Data most commonly used in hydrological studies include:

- □ water level and flow records (this Chapter)
- □ rainfall and climate data (Chapter 4)
- □ soil moisture or groundwater levels (Chapters 8 and 9)
- □ sediment data (Chapter 13)
- □ physical, chemical, and biological measurements (Chapters 12 and 14)

Increasing water scarcity and growing sophistication in analysis are placing an added premium on comprehensive, high-quality data to assist in rapid and economical decision-making.

Hydrological Data

The Hydrological Networks

New Zealand has an extensive network of hydrometric stations for which data are being or have in the past been collected. The Index to Hydrological Sites in New Zealand, the most recent of which is for 1989 (Walter, 1990), provides information on lake level, river level and flow, groundwater, and other types of data collecting stations, such as water temperature measurement sites (Figures 3.1, 3.2).

The first hydrometric station was opened in New Zealand in about 1906. In 1990, there were almost 600 stations at which river level was monitored continuously, and a greater number of closed stations for which data are available for past periods (Table 3.1). The networks of other types of hydrometric stations are similarly extensive. Several types of data are often collected at a given station, particularly water discharge, water temperature, sediment concentrations, and a range of water quality measurements.

Stations are opened and closed to reflect changing information needs, length of record, and finan-

WATERLEVEL RECORDER INDEX - HAWKES BAY

Site no.	Aquacode	River name and site name	Map references: (yard)	(metre)	Area (km²)	Purpose	Recording auth.	Recorder make	Begin date	End date	File arch.	Comment
20101	FL8ICI	Kopuawhara at Railway Br	N116:236951	Y19:308319	54.5	wr,fc	Hbrc	L&S	810429		VR	
20901	F6GTC	Nuhaka at Nuhaka Valley Rd	N116:145960	X19:225329	157.3	wr,ii	Giwrs	Fox	841106		V	
20906	F6UGC	Nuhaka at Hineroa	N106:184101	X19:264457	18.7	ws,wr	Hbrc	Fox	831221	850726	VR	
21301	FP8KHC	Waiatai at Waiatai Rd	N116:859958	X19:964335	14.5	wr,fc	Hbrc	Fox	761215	801015	VR	
21302	FP8MHC	Waiatai at Taits Br	N116:872959	X19:976335	13.2	wr,fc	Hbrc	L&S	790510	850117	VR	Weir
21401	F7IRX	Wairoa at Marumaru	N106:853108	X19:962472	1759	wr,si,mr	Hvwrs	L&S/t, Fox	800215		V	
21409	F72JTN	Waiau at Otoi	N105:480048	W19:620427	513	wr,pi,si	Hvwrs	Fox	680801		V	
21410	F72I2KIC	Waihi at Waihi	N105:564136	W18:699505	49.7	wr,pi,si	Hvwrs	L&S	680801		V	

Notes : Aquacode. A method of identifying sites which enables relative positions to be identified.
Purpose : see Table 3.3.
Recording authority. Hbre: Hawkes Bay Regional Council. Giwrs, Hvwrs: Gisborne and Havelock North Water Resources Survey
File arch. Archive on which data are filed: V: Water Resources Databank; R: regional archive.

Figure 3.1 Typical entries from the Index of Hydrological Recording Sites in New Zealand 1989 (Walter, 1990).

cial constraints. Hydrological data requirements differ with their intended use. Stations operated for water resources assessment or preparation of catchment management plans may be open for relatively short periods - 3 to 5 years - which are sufficient when the emphasis is on water availability at low flow. Stations which provide design or operational data tend to be long-term. Continuous long hydrological records are particularly valuable for studies such as trend analysis, response to catchment and environmental change, and analysis, of extreme events, including floods and droughts, rainfall distributions and catchment response to unusual conditions. In many studies more than one hydrological variable must be monitored. For instance, scientific or engineering design studies often compare catchment rainfall to flow, to both check the validity of the record and to help understand the response of the catchment to rainfalls, particularly extreme events.

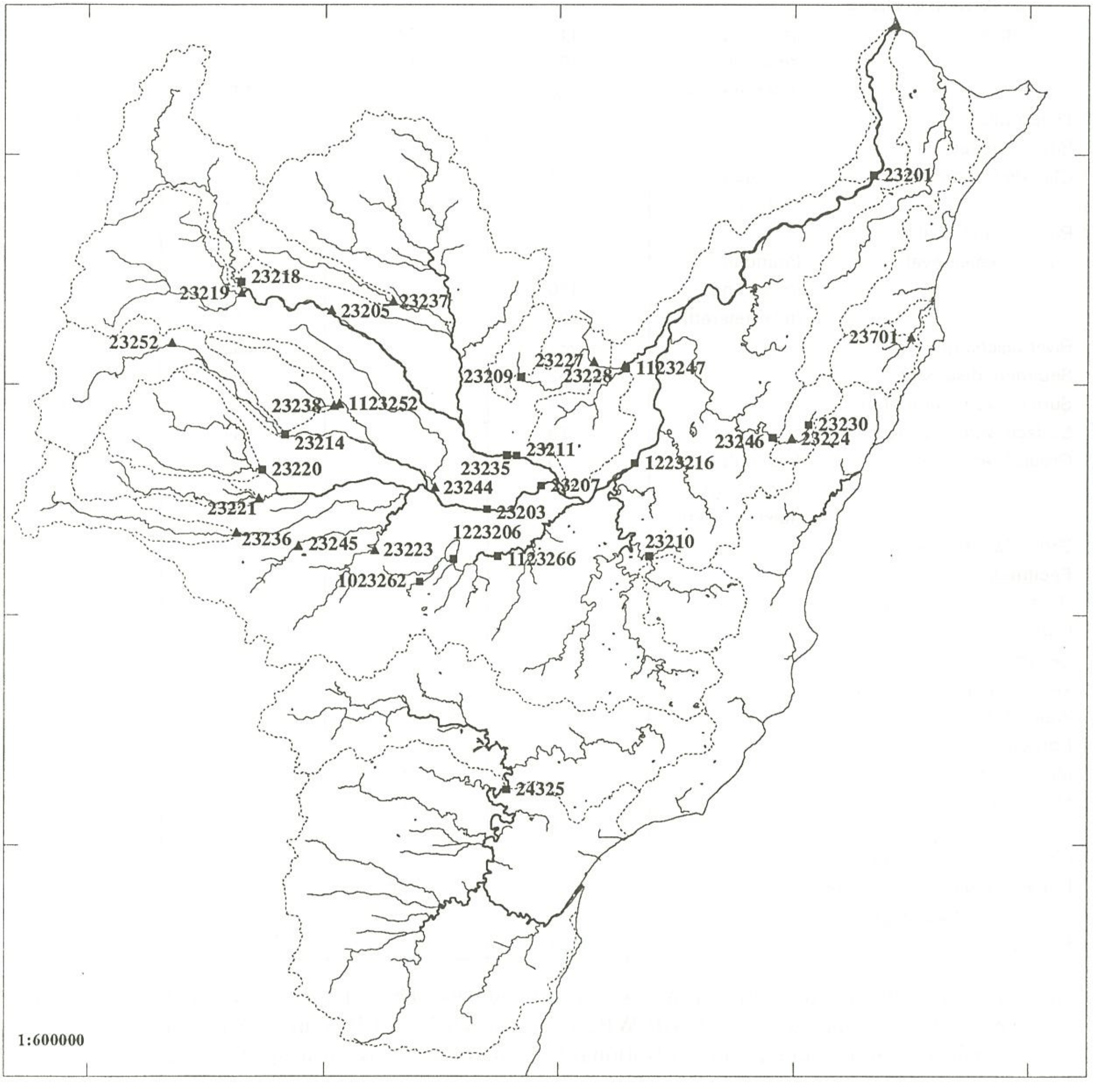
Sources of Hydrological Information

A number of organisations collect hydrological information (Table 3.2). Most hydrological data are or have been collected for specific purposes, such as the design or operation of water resource projects (irrigation, hydro-electricity, flood control), water resources assessment and allocation, and research. The Index of Hydrological Recording Sites in New Zealand (Walter, 1990) identifies the purposes for each station (Figure 3.1, Table 3.3), although the classification probably does not fully reflect the range of uses of the data.

Regional Councils. Regional councils have the statutory responsibility under the Resource Management Act to collect the information required for water resources management. The councils cover the entire country; their boundaries largely coincide with major catchment boundaries, but in some places water is transferred from one council's region to another, generally for hydro generation.

The level of activity in hydrological data collection varies with the level of economic development of the regions and the importance of water resources - or, more often - their scarcity. Some regional councils emphasise flood warning and control, so their data collection systems are directed to this end rather than to resource assessment. Nevertheless, all regional councils maintain archives of hydrological data, which may contain large volumes of data.

All activities (other than domestic use, stock



Hawkes Bay B Waterlevel Recorders

Figure 3.2 Example map from the Index to Hydrological Recording Sites in New Zealand 1989 - Hawkes Bay waterlevel recorders (Walter, 1990, p37).

Evaluation Element		DSIR WRS*	NZ Met Service	Regional Councils*	Total
Data Collection Stations					
Precipitation	(manual)	113	948	272	1333
	(recording)	101	120	128	349
	(telemetered)	28		137	154
Evaporation			58	3	61
Snow courses		12		0	12
Climate stations	(manual)	1	181	1	183
	(automatic)		47	1	48
Precipitation quality				0	0
Surface water level	(manual)			270	270
	(recording)	170		226	396
	(telemetered)	129		135	264
River discharge		261		332	593
Sediment discharge		102		67	169
Surface water temperature		261		173	434
Surface water quality		78		234	312
Groundwater level	(manual)	5		1325	1330
	(recording)	14		120	134
	(telemetered)			4	4
Groundwater quality				132	132
Facilities					
Repair and Maintenance		1	1	2	
Current meter rating		1		0	1
Sediment laboratory		6		3	9
Water chemistry laboratory		2		7	9
Well drilling sets			1	1	
Personnel					
Met station observers			87	87	
Met technicians			160	0	160
Hydrology technicians		65		45	110
Computer staff/data process		2		24	26
Professional meteorologists			80	0	80
Professional hydrologists		1		22	23
Supervising staff		1		6.3	7.3

*Note: data available for only nine of the 14 regional councils. Those included probably account for 80% of all the total activity. DSIR WRS is the former Water Resources Survey of DSIR, now Environmental Data Division of National Institute of Water & Atmosperic Research.

Table 3.1 Data Collection Installations, Support Facilities, and Staff (Mosley, 1990)

water, or fire-fighting) which modify natural waters or watercourses - abstraction, discharges, gravel extraction, channel modification - require water right or permit from the relevant regional council. Such rights may specify that the abstraction be monitored. Data are collected for most, if not all, medium to large industrial projects (power generation; water abstraction for domestic and industrial use;discharges from waste treatment plants). Major irrigation schemes are also monitored, although the emphasis is often on critical conditions, such as the threshold low flows at

Regional Councils

Auckland Regional Council, Private Bag 92012, Auckland
Bay of Plenty Regional Council, PO Box 364, Whakatane
Canterbury Regional Council, PO Box 345, Christchurch
Gisborne District Council, PO Box 747, Gisborne
Hawkes Bay Regional Council, Private Bag 6006, Napier
Manawatu-Wanganui Regional Council, Private Bag 11025, Palmerston North
Marlborough District Council, PO Box 443, Blenheim
Northland Regional Council, Private Bag 9021, Whangarei
Otago Regional Council, Private Bag, Dunedin
Southland Regional Council, Private Bag 90116, Invercargill
Taranaki Regional Council, Private Bag 7713, Stratford
Tasman District Council, Private Bag 4, Richmond 7002
Waikato Regional Council, PO Box 4010, Hamilton
Wellington Regional Council, Po Box 11646, Wellington
West Coast Regional Council, PO Box 66, Greymouth

Central government and State-owned enterprises

Landcare Research Ltd, PO Box 40, Lincoln
Environmental Data, National Institute of Water & Atmospheric Research, PO Box 3047, Wellington
Meteorological Service of NZ, PO Box 722, Wellington
Project Services, Works Consultancy Services, PO Box 12-447, Wellington
Fuel Resource Group, Electricity Corporation, PO Box 930, Wellington

Table 3.2 Sources of Hydrological Information

which abstraction must be reduced or cease. There are many hundreds of minor users - small, single farm irrigation systems, wells for single dwellings, waste discharges from dairy farm milking sheds - which are not monitored.

Central Government. Agencies active in hydrological data collection are the National Institute of Water & Atmospheric Research, Landcare Research, and the Institute of Geological and Nuclear Sciences. Other Crown Research Institutes also collect data, primarily for research (eg small catchment experiments in forest areas; climate stations at experimental farms).

Environmental Data, a division of the National Institute of Water & Atmospheric Research, operates the only nationwide data collection programme for surface water, designated the National Hydrometric Reference Network (NHRN). The Network includes, in each hydrologic region, at least one river flow gauging station per 5000 km^2, distributed evenly across all sizes of catchments (with all catchments $>$800 km^2 at the sea having a station as close to the sea as practicable). It provides an information base for research at the national scale, and many stations are former representative, regional, or major river basins (Ministry of Works, 1970). The network also includes some stations originally installed for operational purposes such as power investigation, and selected to achieve the desired even density throughout the country.

A National Water Quality Network was estab-

bi	Bridging investigation	pt	Power investigation-thermal
fc	Flood control investigation (or river control investigation)	rg	Regional catchment
fw	Flood warning	rs	Research area
ii	Irrigation investigation	si	Sediment investigation
ll	Lake levels	tr	Tidal recording
lu	Land use catchment	ug	Underground water investigation
mr	Major river	ur	Urban site
nhrn	National Hydrometric Reference Network	wq	Water quality monitoring
pi	Power investigation	wr	Water resources
pl	Local power board investigation	ws	Water supply investigation
po	Power operation		

Table 3.3 Purposes for which Hydrological Stations may be operated in New Zealand (Walter, 1990)

lished in 1989, consisting of 78 river stations at which data are collected to specifications established by the Ecosystems Division of the National Institute of Water & Atmospheric Research. About thirty lake stations were added in 1992.

Environmental Data also operates a National Sediment Network for research, consisting of subnetworks as follows:

- □ national flood-period observations (about 100 sites)
- □ coastal sediment yield (about 15 additional sites)
- □ national regime (17 representative sites)
- □ environmental regime - national importance (8 sites).

Private Enterprise and State-owned Enterprises.

Because of the extensive requirements for monitoring abstractions, discharges etc, and for environmental impact reporting, many enterprises fund hydrological data collectoin by contracting regional councils, the CRIs, and others to carry out the work.

A major acquirer of hydrological data is Electricity Corporation, which has two parallel systems - SCADA, an in-house system for monitoring power station throughput, lake levels etc for controlling the generating system, and a hydrometric survey operated on contract by the National Institute of Water & Atmospheric Research in conjunction with its own work. The Corporation operates hydropower stations throughout the country, and also funds gauging stations for hydro potential assessment. Works Project Services carries out hydrological work on behalf of Electricity Corporation, and maintains archives of information and reports.

Data may be sought from the central and regional government agencies referred to in Table 3.2, although a charge may be made for the cost of data retrieval, and any necessary processing. In general, data collected with public funds are accessible by the public, although a commercial user may be charged a royalty in addition to data retrieval costs.

Adequacy of Coverage of Hydrological Networks

Average densities of stations, as recommended by the WMO (1981), have little relevance to New Zealand. Densities vary widely between regions and catchments, and most stations have one or more specific purposes - that is, there is no general-purpose national network. Nevertheless a network of national extent exists in practice, because data collected by the various organisations are in general complementary and readily available.

Hydrologic parameters are very variable in New Zealand, because of complex topography and steep climatic and hydrological gradients. As a result, achievement of theoretically desirable levels of data accuracy nationwide would require very much denser networks than actually exist. However, the greatest variability is found in mountain areas of little economic importance, so increased numbers of monitoring stations could not be justified.

Hydrological Instrumentation and Procedures

The recording of discharge of streams and rivers normally involves (1) obtaining a continuous record of water levels, or stage above a datum; (2) establishing the relationship between water level and discharge (the rating curve); and (3) transforming the record of stage into a record of discharge. The site at which water levels are recorded and discharge is measured is commonly called a gauging station.

Collection of lake level or river level data (e.g. for flood-warning) requires only step (1), although in fact many lake level records are used to calculate outflow rates, for example the Buller River as it flows from Lake Rotoiti. In this case, steps (2) and (3) also are required.

Other types of measurement commonly undertaken at gauging stations include suspended sediment concentrations, water temperatures, and other aspects of physical, chemical, and biological water quality (chapters 12, 13).

Hydrological practice in New Zealand is aligned with that of the US Geological Survey (Rantz et al 1983), and as recommended by the

World Meteorological Organisation (WMO 1981, 1988) and International Standards Organisation (1983).

Standards for Hydrological Data

The starting point for defining hydrological practice is data quality. Standards for data to be used for scientific purposes have been defined by Environmental Data (Table 3.4), and are applied to its stations, to those operated for Electricity Corporation, and to an increasing number of stations operated by regional councils. However, many stations - particularly those installed primarily for flood warning - need not usually operate to such stringent and expensive specifications.

Consistent application of defined standards requires that comprehensive manuals of practice are available. Those which have been adopted by Environmental Data and many regional councils, which include several prepared in New Zealand, are listed in Table 3.5.

Selection of a Site for a Gauging Station

The most important requirement for the location of a gauging station is that the downstream hydraulic controls, which control the water level at the station and determine the relationship between water level and discharge are stable, and sensitive to changes in discharge. That is, the rating curve should not change over time, and a measurable change in discharge should be matched by a measurable change in level. Often, conditions may change with flow, and the feature which controls water level at low flow may be drowned out at higher flows as the influence of a downstream control extends upstream (Figure 3.3.).

Hydraulic controls are of several types (Arnold et al, 1988, p 6-14) :

- **a weir-type control** is probably the most common - a rock bar or gravel bar at the head of a riffle or rapid controls the water levels in the upstream pool;
- **constrictions or bends** can control water levels, particularly at higher flows;
- **friction control** occurs in long straight uniform channels, where the water depth is determined by channel shape and roughness;
- **artificial controls** such as concrete sills across the streambed are often installed to stabilise the channel bed, and at low flows act as either weir-type or friction controls.

In many smaller streams, weirs or flumes have been installed where satisfactory stable natural controls were unavailable. They have predetermined rating curves (Ackers et al, 1978), but

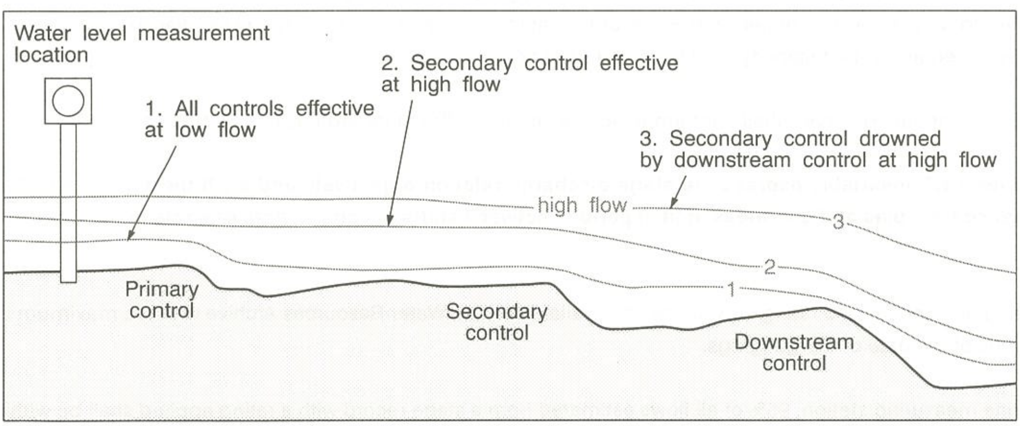


Figure 3.3 The influence of downstream controls on water level.

Table 3.4 Standards adopted by the Environmental Data Division of NIWAR, (formerly Department of Scientific & Industrial Research)

To provide hydrometeorological information of sufficient accuracy to meet the objectives of the National Hydrological Reference Network, the following standards have been specified:

NOTE: Incidents outside the control of the Environmental Data Division (eg. vandalism or major storms), may require that these standards are temporarily forgone for the period of affected record.

Water Level (River, Lake and Groundwater) and River Flow Sites

Measurement of stage

(a) Installed equipment and operating procedures shall be such as to ensure that 95% of instantaneous measurements are within ±3 mm or ±10 mm of the water-level above the sensing device, depending on site instrumentation; specifications in clause 7 of ISO Standard ISO 4373-1979(E) should also be met.

(b) Instantaneous values shall be available on the Water Resources Archive within a maximum of six months of their being recorded.

(c) At any one measuring station, there shall be not more than 2% (approx. 7 days) missing record in a given calendar year, and not more than one calendar year in ten shall have any missing record.

(d) Field practice shall conform to standards specified in the Water Resources Survey Hydrologists' Field Manual (DSIR, 1988)

Measurement of discharge, and rating curve construction

(a) Flow gaugings shall conform to the appropriate ISO standards as outlined in ISO Handbook 16 (ISO, 1983), and documented in the Water Resources Survey Field Hydrologists' Manual; in any case, 95% of discharges shall be measured to an accuracy of better than ± 8% of the rated value, and at a frequency specified by reference to flood event frequency, bed stability, and historical evidence.

(b) Stage discharge rating curves shall conform to specifications of ISO Standard ISO1100/2-1982, clause 7.1:

the curves shall invariably express the stage-discharge relation objectively and shall therefore be tested for absence from bias and goodness of fit in periods between shifts of control, and for shifts in control (ISO, 1983).

(c) Flow gaugings and revised rating curves shall be available on the Water Resources Archive within a maximum of six months of the date of the gaugings.

(d) At any one measuring station, 95% of all flows estimated from a stage record with a rating applied shall be within ± 8% of the actual values.

Water Quality Measurements

Temperature

(a) Field practice shall conform to standards specified in the Water Resources Survey Hydrologists' Field Manual (DSIR, 1991).

(b) All data collected and archived shall be within ± 0.5°C of a calibrated (TELARC registered) check thermometer'

(c) Data shall be available on the Water Resources Archive within a maximum of six months of being recorded.

Sediment concentration

(a) Field and laboratory practices shall conform to standards specified in the Water Resources Survey Suspended Sediment Manual (DSIR, 1991).

(b) Data shall be available on the Water Resources Archive within a maximum of six months of their being recorded.

Table 3.5 Manuals Of Practice Commonly Used In New Zealand

1. Hydrology Manuals (1990), consisting of :

   WRS Hydrologists' Field Manual (1988), P E Arnold, G Holland, A I McKerchar, and W R Soutter
   WRS Office Practice Manual (1992), D A McMillan
   WRS Telemetry Users' Manual (1989), M B Thomas and M J Butler
   GAUGE Users' Manual (1990), M B Thomas
   RHYHABSIM Computer Manual (1989), I G Jowett
2. TIDEDA Reference Manual (1991), M W Rodgers and S M Thompson
3. Drawing and Checking Stage/discharge Rating Curves (1987), A I McKerchar and R D Henderson
4. Index to Hydrological Recording Sites in New Zealand (1989), K M Walter
5. COMMENT User's Manual (1990), M L Wilmer
6. Hydrologist's Safety Manual (1984), J K Fenwick and R J Curry
7. Measurement and Computation of Streamflow: Volume 1. Measurement of Stage and Discharge (1982), S E Rantz and others
8. Measurement and Computation of Streamflow: Volume 2. Computation of Discharge (1982), S E Rantz and others
9. Field Methods for Measurement of Fluvial Sediment (1970), H P Guy and V W Norman
10. AQUITEL 2 User's Manual (1984), Harding Signals Ltd
11. MS-DOS Users' Guide (1986), Microsoft Corporation
12. Policy on the Management of Water Resources Archive (1991), DSIR
13. Quality Manual (1991), DSIR
14. A Procedure for Constant-rate Injection Salt Dilution Gauging (1989), D E Johnstone

A

should be calibrated by gaugings in the field to confirm the continued applicability of the rating curve, which may change in response to vegetation growth, structural settlement, or altered approach conditions.

In a natural channel, the rating curve is established by measuring discharge and water level simultaneously, across the observed range of flow conditions. Manual discharge measurement using a current meter requires a location on a straight section of channel where the streamlines of flow are approximately parallel, and flow velocities are in the range that can be accurately measured with a current meter (0.03 to 6 m/s). The location need not be the same as that chosen for the water level recorder, but the discharges must be exactly the same.

Water Level Measurement

Equipment for measuring and recording water level in New Zealand is of several types (Arnold et al, 1988, p 16-25) :

- ☐ float in a stilling well, connected by a counter-weighted tape to a chart recorder, punched-tape recorder, or shaft encoder which sends an electronic signal to a datalogger (Figure 3.4);

B

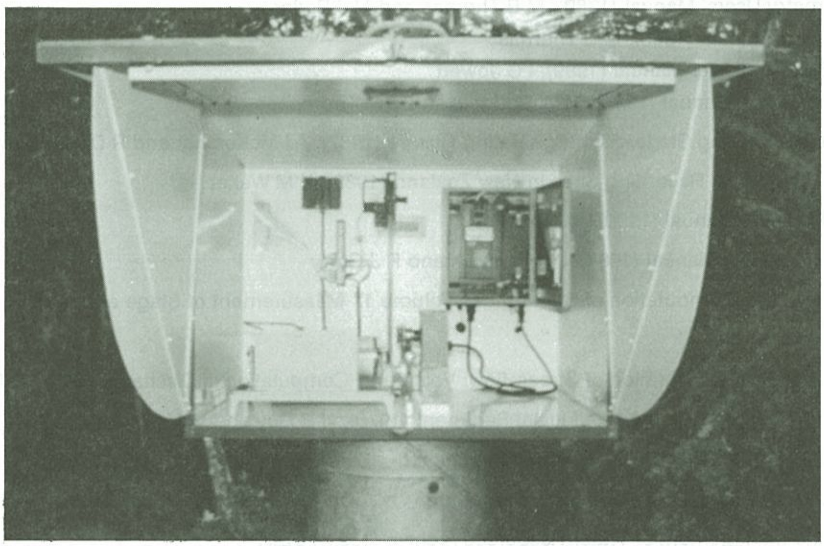
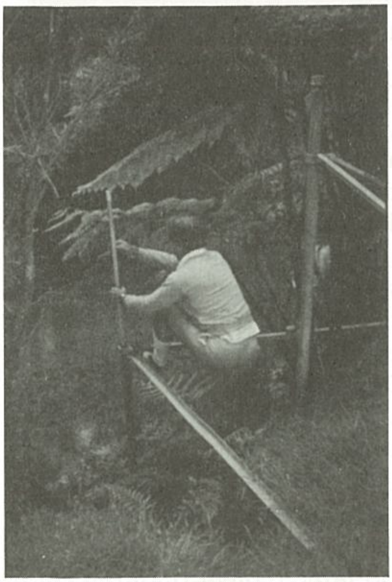


Figure 3.4 Typical water level measurement stations. **A.** The stilling well and recorder housing on the Cropp River. **B.** Instrumentation inside the station at Wharauroa Stream.

A

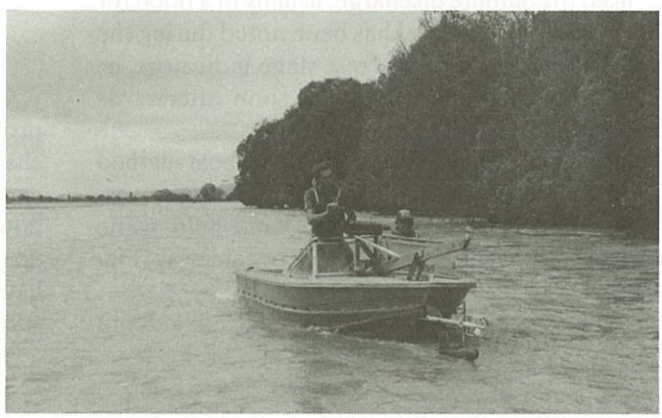
B



C

D

Figure 3.5 Methods of gauging discharge.
A. Dilution. **B.** Current meter gauging on the Onyx River, Antarctica. **C.** Crest-stage recorder on the Awana Stream, Great Barrier Island, used to establish a high flow rating curve for an inaccessible station. **D.** Cableway current meter gauging on the Haast River. **E.** Jetboat gauging.



E

- pressure bulb and chart recorder;
- pressure transducer, which sends an electronic signal to a datalogger;
- gas purge, in which the pressure differential above a gas bubbler outlet is sensed by a pressure transducer or beam balance mechanism.

Electronic dataloggers avoid the expense associated with digitising traces on paper charts or processing punched-tapes, and allow integrated electronic data processing. Many stations are equipped with telemetry, which transmits data back to base via radio or telephone. Transmissions can be in response to alarm levels (eg for flood warning), in response to requests from the base station, or automatically (e.g., regular daily transmissions to the base station archive).

Discharge Measurement

A wide range of methods is used for measuring discharge in New Zealand, including :

- volumetric, in which the time required to fill a calibrated container which receives all the flow (usually at the outfall of a weir) is measured. This method is suitable only for flows less than 0.2 m^3/s;
- dilution, in which the change in concentration of a liquid (usually dye or brine) introduced into the natural flow can be used to calculate the discharge (Figure 3.5A);
- velocity-area, in which the velocity and depth at a number of points across the channel are measured, and used to compute discharge. Velocity may be measured with a current meter, timed float, or velocity head rod (Figure 3.5B);
- slope-area, in which the Manning equation is used to compute discharge, usually of a flood for which the water level has been noted during the event, measured with crest stage indicators, or inferred from debris marks soon afterwards (Figure 3.5C).

Ultra-sonic gauging and the moving boat method are little used in New Zealand; most discharge measurements use current meters, hand-held while wading, or suspnded from a bridge or cableway (Figure 3.5D), or from a boat (Figure 3.5E).

With current meters, velocity and water depth are measured at a number of locations, or verticals, spaced across the cross-section (Arnold et al 1988, p 40-69). The verticals are not necessarily equally spaced, but are chosen so that obvious variations in velocity or bed shape are measured, and no vertical represents more than 10% of the flow. Normally, measurements are made at 20-25 verticals.

Water velocity in an open channel is at a maximum near the surface and approaches zero at the bed (see Figure 14.1). The mean velocity at a vertical is typically at 0.6 of the depth from the surface, and it is common practice, in water less then 0.75 mm deep, to take a single velocity measurement at this point. In deeper water, velocities are measured at 0.2 and 0.8 of the depth and averaged. Where there is reason to think that the vertical velocity profile is distorted from its normal semi-logarithmic shape, more measurements will be made, for example at 0.2, 0.6, and 0.8 of the depth.

Current meters contain a rotating element whose speed of rotation is proportional to the water velocity. Those used in New Zealand have a propeller (e.g., the Ott type) or a vertically mounted rotating cup (the Price type) (Figure 3.5B). Calibration tables or equations are needed to convert speed of rotation of the current meter element to water velocity. The calibrations are developed by towing the current meter at a series of known velocities through still water in a rating tank. New Zealand's only rating tank is at the Environmental Data Support Centre, Kainga.

The observations of depth, velocity, and spacing of verticals are used to compute the incremental discharges across the section, and thence the total discharge, using the equation (Figure 3.6) :

$$q_i = V_i\,[(b_{i+1}-b_i)/2 + (b_i-b_{i-1})/2]d_i \qquad (3.1)$$

$$= V_i\,[(b_{i+1}-b_{-1})/2d_i$$

There are many complications to this simple procedure, for example where the discharge is changing rapidly during the gauging, or where a current meter suspended from a cable is displaced downstream by fast flowing water and therefore does not hang vertically. Calculation methods have been developed for such cases (Rantz et al, 1983; Arnold et al, 1988).

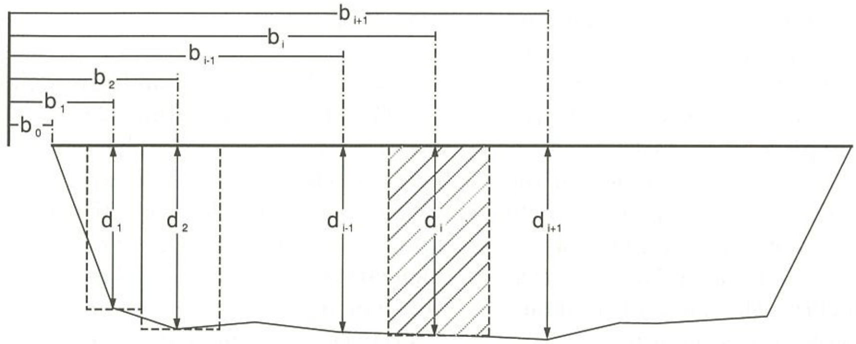


Figure 3.6 Delineation of a cross-section for measurement of discharge by the velocity-area method.

Stage-discharge Rating Curves

The rating curve (McKerchar and Henderson, 1987) permits conversion of the stage record into a discharge record (Figure 3.7). At many stations in New Zealand, there are periodic changes in the flow controls, caused by scour or aggradation during floods, weed growth during prolonged low flow, gravel extractions downstream, and so forth. A sequence of rating curves, constructed from regular gaugings, may be required to attain the standards defined in Table 3.5.

In a long uniform channel, a rating curve has the form

$$Q \;=\; C\,(h-a)^N \qquad (3.2)$$

where Q is discharge, C and N are constants, h is stage, and a is stage at which discharge is zero. Theoretical values of N for rectangular and parabolic channels are 1.67 and 2.17 respectively, assuming width is greater than 20 times depth. The value for a triangular channel is 2.67. Equation 3.2 also applies to weirs.

Equation 3.2 can be used to fit a rating-curve to a series of discharge measurements. This method is objective and provides an error of estimate, but in simple form requires that one control acts over the whole range of flows. An additional control requires a second curve to be fitted, and the point of transition to be identified.

Rating curves are normally prepared graphically, by plotting gaugings on an arithmetic stage-discharge plot, and drawing in the curve by eye. Knowledge of the cross-section's behaviour, such as the effect of multiple controls, the response to aggradation and scour, whether rating curves are parallel or converge on a single relationship at high

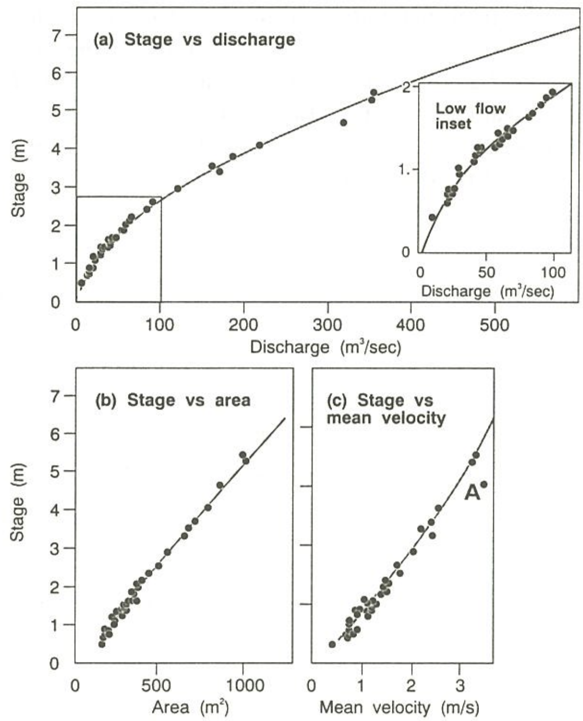


Figure 3.7 Stage-discharge rating curve for the Grey River at Dobson, Westland.

flows, is often needed for an accurate result, and analytical fitting methods (Herschy, 1985) have not to date been used much in New Zealand.

Commonly, stage versus cross-sectional area and mean velocity is plotted, to assist in interpretation, extrapolation and construction of the stage-discharge rating curve. They may also help identify errors that are not apparent in the stage-discharge plot (Figure 3.7), and delineate variations from a smooth curve which are due to multiple controls and complex cross-sections.

Often, a rating curve must be extrapolated beyond the range of gauged discharges, because a gauging station has never been visited during the highest flows (telemetry is useful in alerting the hydrologist and enabling visits to be arranged for high flow gaugings). For stable cross-sections, the curve can be extrapolated by extending the stage-area and stage-velocity curves, and taking the velocity-area product at a given stage to estimate discharge.

Data Processing

In New Zealand, almost all hydrological agencies use the TIDEDA (TIme DEpendent DAta) software for processing, archiving, and presenting their data (Rodgers and Thompson, 1991). Other software is used for dealing with various aspects of data processing, such as TELSYS (operating telemetry systems) and GAUGE (calculating flows from gauging data).

TIDEDA presents data in a range of formats, including :

- ☐ tabulation of daily mean discharges, organised in a variety of ways;
- ☐ hydrographs and other time series plots (Figure 3.8);
- ☐ cumulative frequency distribution curves;
- ☐ cumulative mass curves;
- ☐ rating curves;
- ☐ tabulations of extreme values.

It has a simulation capability which can be used

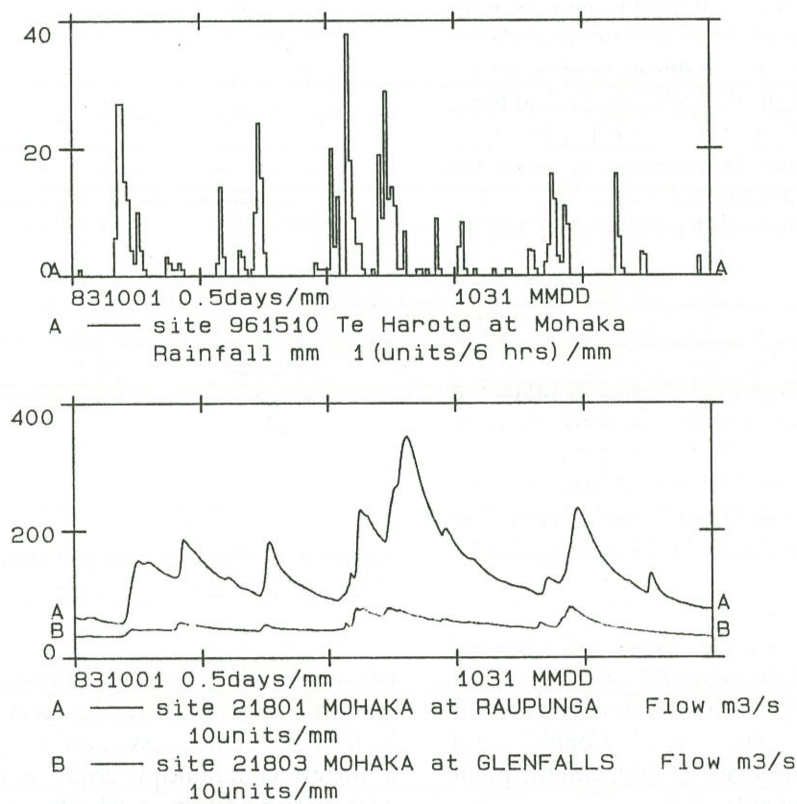


Figure 3.8 TIDEDA-generated plot of rainfall and flow hydrographics (Rodgers and Thompson, 1991, p84).

for sophisticated data manipulation and analysis, for example to study the effect of different flow allocation regimes on residual flows in a river.

Procedures for data processing are closely specified by Environmental Data in order to ensure a standard product and to avoid any chance of data corruption (McMillan, 1992). This procedure, or variants thereon, are in general use in New Zealand.

A critical, but often overlooked, element of hydrological data acquisition is the need for quality assurance, the assurance that data meet defined standards and are usable for the intended purpose (Mosley and McKerchar, 1989). Environmental Data has implemented a comprehensive quality assurance programme which is accredited by Telarc, New Zealand's third-party testing and standards accreditation organisation. The programme meets the standards specified in International Standard ISO-9002. It is based on a Quality Manual (McMillan, 1992) and a comprehensive set of manuals which define all aspects of hydrological data acquisition, processing, archiving, and instrumentation. Many regional councils are implementing similar, but generally less formalised, quality assurance procedures.

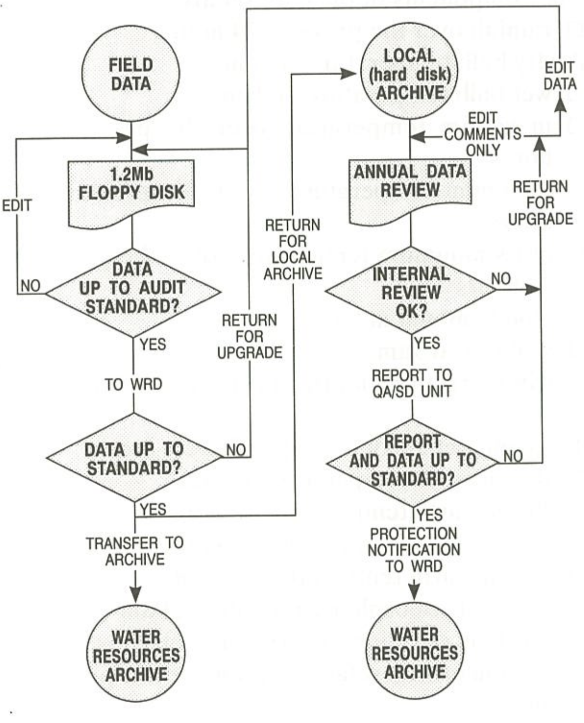


Figure 3.9 The sequence of quality assurance checks followed by data archived on the Water Resources Archive (McMillan 1992, p2-26).

A quality assurance programme requires careful checking and auditing of the product; Figure 3.9 shows the sequence of data checking followed by Environmental Data staff in updating data to the Water Resources Archive. Such work requires highly trained technicians, and hydrological staff in New Zealand undergo extensive training on the job and through short courses, usually for a minimum of two years.

Meteorological Data

The Meteorological Networks

Introduction

The Meteorological Service of NZ Ltd (MetService - formerly the NZ Meteorological Service) and the National Institute of Water and Atmospheric Research operate various meteorological and climatological networks designed for a range of functions, but which furnish data that are very useful and often essential in hydrological studies.

The synoptic weather network

The synoptic weather network (SYNOP) comprises stations whose prime purpose is for the production of synoptic weather charts for real-time weather forecasting. The number and location of stations has varied greatly over the years, but in 1992 the network had 86 stations (Figure 3.10). SYNOP stations record weather parameters at least once a day, and at some stations up to eight times a day. The reporting times, which are standardised throughout the world, are 0000, 0300, 0600, 0900 etc to 2100 UTC (Universal Time). There are more reports in day-time than night-time, so that in New Zealand most reports tend to be at the reporting times 1800, 2100, 0000, 0300 and 0600 UTC (corresponding to 6am, 9am, noon, 3pm and 6pm NZ Standard Time).

Items of hydrological interest recorded by the SYNOP network include wind velocity, cloud cover, rainfall, temperature and humidity. Other items recorded such as atmospheric pressure, and present and past weather conditions are occasionally used by hydrologists.

The hourly weather network

Sixteen stations sited mainly at airports provide hourly observations to provide the real-time weather data needed for aviation, and the weather reports produced are called METARS (meteorological aerodrome reports). Observations are similar to those made at synoptic weather stations, and indeed the network is largely a subset of the synoptic network. There are more reports during the day than at night; six stations provide a full 24-hour coverage. As with the synoptic network, the wind, temperature, humidity, and rainfall information can be useful for hydrological studies.

The automatic weather station (AWS) network

The 47 AWS sites fulfil the functions of synoptic and hourly weather networks. Generally they are interrogated each hour and the synoptic (3-hourly) reports are archived. For ten stations, the most modern ones equipped with visibility and cloud sensors, hourly reports are also archived. The automatic weather station network (Figure 3.10) will progressively expand as it becomes more cost effective way of gathering weather data compared with manual synoptic and climatological stations.

The climatological network

This network exists primarily for the study of New Zealand's climate (Figure 3.11). It has been operating since the eighteen fifties, and the number of stations in the network has increased since then, although with some periods of declining numbers. A very substantial decline has occurred in recent times, from 380 stations in 1988 to just over 200 in 1992.

In the climatological network, meteorological observations are recorded once a day at 9am local time. The network historically was not for real-time measurement, and the daily 9am recordings were sent in for processing and archiving only once a month. Since 1984 the network has been progressively converted to a real-time data collection system, called the **daily climate** network (DLYCLI). It presently comprises 75 stations that are a mixture of synoptic and climatological stations. These stations are equipped with the weather observers terminal system (WOTS) to send their climatological data to Wellington via Pacnet. As resources become available more climatological stations will be converted to real-time reporting.

The climatological stations record a set of climatic parameters which consists of a core set and an expanded set. All stations record the core set. The expanded set requires more instrumentation and is reported by a subset of the 200 climatological stations.

The components of the core set are:

- ☐ rainfall over the previous 24 hours
- ☐ dry bulb temperature at 9am
- ☐ wet bulb temperature at 9am
- ☐ maximum temperature over the previous 24 hours
- ☐ minimum temperature over the previous 24 hours
- ☐ grass minimum temperature over the previous 24 hours
- ☐ cloud amount at 9am
- ☐ visibility at 9am
- ☐ wind direction and Beaufort force at 9am

The components of the expanded set are:

- ☐ 10 cm earth temperature at 9am
- ☐ 20 cm earth temperature at 9am
- ☐ 30 cm earth temperature at 9am
- ☐ 100 cm earth temperature at 9am
- ☐ sea level atmospheric pressure at 9am
- ☐ sunshine duration over the previous day
- ☐ amount of total solar radiation over the previous day
- ☐ wind run over the previous 24 hours
- ☐ maximum wind gust speed and time over the previous day
- ☐ evaporation over the previous 24 hours

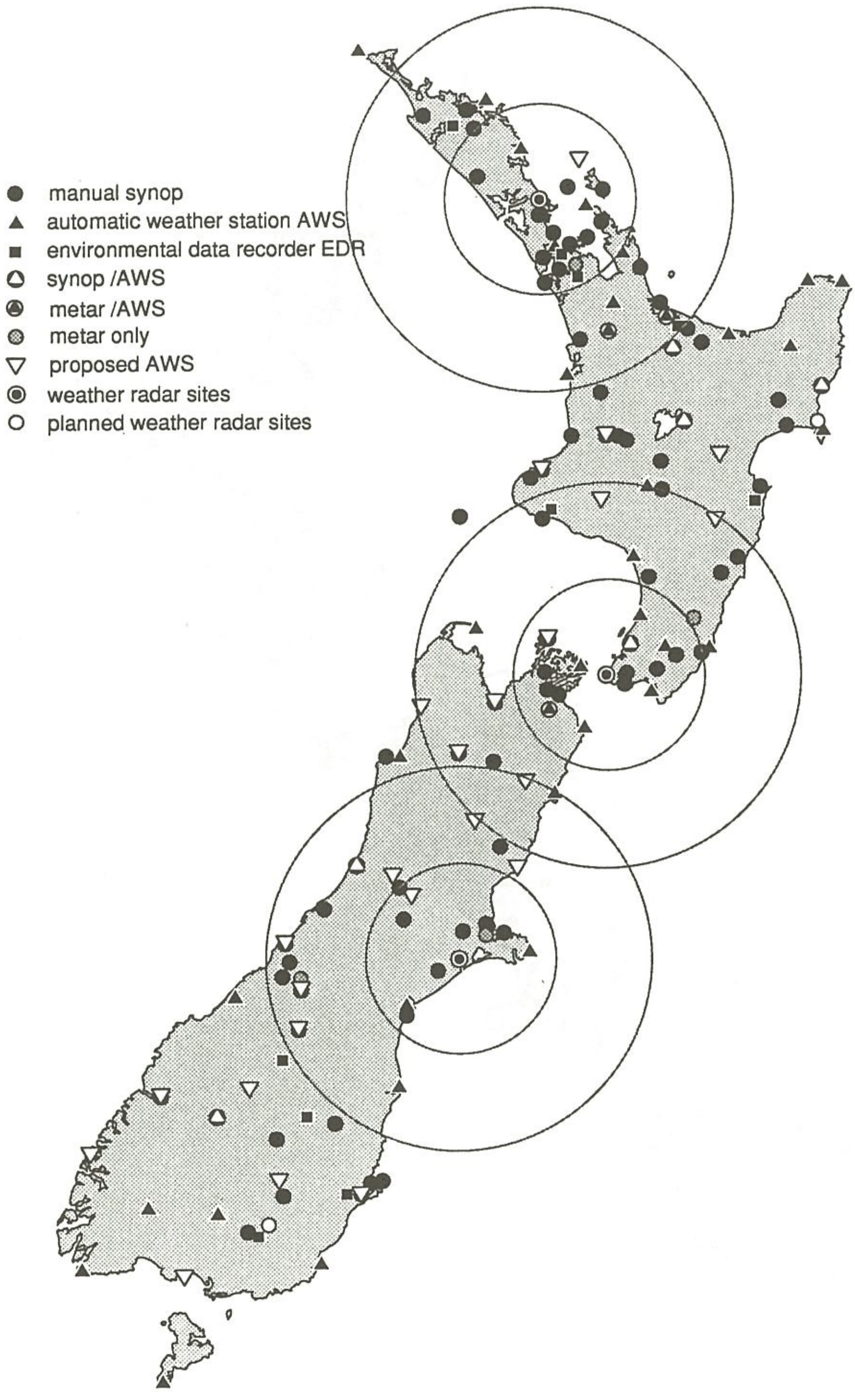


Figure 3.10 The SYNOP, METAR, and AWS meteorological networks at January 1992. The location and coverage are shown of weather radar stations in New Zealand. The planned locations of the Southland and central North Island radars may change.

Figure 3.11 The climatological station network.

- □ note of the occurrence in the previous 24 hours of: gale, hail, lightning, thunder, fog, dew, snow
- □ depth of fresh snow at 9am
- □ total snow depth at 9am
- □ weather sequence over the previous 24 hours

Other factors, such as relative humidity, dew point, and evapo-transpiration are derived from these measurements.

Many climatological stations have chart recorders for recording temperatures, rainfall, pressure, sunshine, radiation and other items. Some of the autographic records from these chart recorders have been processed, often on an hourly basis. Rainfall charts are progressively being processed to produce a digital record of the complete chart record. At the same time chart recorders are being progressively replaced with digital recorders.

The rainfall network

This is the simplest network, consisting largely of stations where daily rainfall is recorded just once a day at 9am local time. The value recorded is the daily rainfall and it is assigned to the day prior to the recording day. The rainfall stations furnish monthly returns. The average life of a station in this network is about 18 years, with records ranging from 1 to 2 years to over 100 years. There have been about 1800 stations in the network, and of these some 670 were operating in 1992. There is a sub-network of about 100 chart recording gauges, for each of which maximum monthly rainfalls of various durations from 10 minutes to 72 hours have been extracted and filed. Some long records of these maximum rainfalls are available. The rainfall charts are currently digitised for computer storage.

The weather surveillance radar network

In 1992 the weather radar network was still being established. The planned network consists of five S-band (5cm) radars (Figure 3.10). The radar have an effective range of around 240 km. Because of earth curvature the effective range for quantitative precipitation is about 120 km. Readers requiring a brief background on the operation of weather radar may consult Collier (1989) or Bogush (1989).

There are two major limitations on the use of these weather radars, apart from the obvious one that significant parts of New Zealand will not be covered. The first is ground clutter, which is a serious problem in a mountainous country like New Zealand. Mountains restrict the areal coverage of the radar by intercepting the radar beam. The Canterbury radar, for example, is severely restricted in its view to the west by the foothills and the Southern Alps. The Wellington radar sends back images from waves in and around Cook Strait when the surface winds are strong. Clutter can be attended to in the radar data analysis software, but users still need to be aware of its existence.

The second problem is that the five centimetre radar beam is attenuated in heavy rain, giving inaccurate rainfall rates. For example, for a 20 km diameter shower whose central rainfall rate is 100 mm/hr and whose rainfall rate gradient out from the centre is 10 mm/hr/km, the radar will detect only about 50 mm/hr for the central rate. This attenuation severely reduces the radar's ability to detect areas of very heavy rainfall, especially if such rainfall lies behind other moderate or heavy rainfall. Users of radar imagery need to be aware of this problem.

Satellite Coverage

There are two types of satellite coverage; polar orbiter satellites and geostationary satellites. Both systems have operated for twenty years or more but have been progressively improved over that time.

There are a series of about six weather satellites, evenly spaced, in geostationary orbit at an altitude of 36,000 km over the equator. New Zealand falls currently under the 'view' of a Japanese satellite located at longitude 140 degrees east, and a United States satellite at 135 degrees west. Images in both the infra-red and visible wavelengths are received from these satellites each three hours. They have a resolution of about four kilometres at New Zealand latitudes. The images are not currently permanently archived by Met Service NZ Ltd but permanent archives are being developed by Geophysics Institute, Victoria University, and possibly by other groups.

The polar orbiting satellites are at a lower orbit and offer higher (2 km) resolution images. These satellites have a life of typically 3 to 5 years. They are operated by the National Oceanic and Atmospheric Administration in the United States. Beside providing visible images their imagery allows for processed fields to be produced for ground surface moisture and temperature.

Permanent archiving facilities are being developed for the polar orbiting satellite images. In the meantime, copies of recent images are always available.

Instrumentation

Rainfall Measurement

Rainfall and other meteorological measurements follow the instructions and standards established by the WMO (1981, 1988) and further described in the British Meteorological Services Observers Handbook (1962) and the NZ Meteorological Services Internal Manual of Instructions.

Several types of rain gauges are used in New Zealand. They fall into three categories: manual gauges read daily, manual storage gauges read weekly or monthly, and recording gauges. Accurate measurement of rainfall is not easy, and a vital consideration is the gauge's exposure. Generally the accuracy sought in a rainfall measurement is 0.1 mm, but storage gauges may be less accurate. The rainfall caught in a standard rain gauge is strongly influenced by the airflow around the gauge. Airflow, in turn, is influenced by the size of the gauge, the height of its orifice above the surrounding surface, the presence of objects in the vicinity of the gauge, and the velocity of the prevailing wind at the gauge. Strong winds tend to reduce the catch of rainfall since the drops tend to be blown past the orifice.

The recommended standard exposure is to have the gauge orifice 30 cm above the ground surface - preferably a horizontal, closely cut grass surface - with any nearby objects at a distance of at least four times their height. This reduces the likelihood of turbulent wind flows around the gauge producing erroneous catch values. To improve exposure a number of techniques are sometimes employed. These include slatted shields around the gauge, or a 30-cm high turf wall built around the gauge at a distance of about 1.5 metres, or anti-splash mats around the gauge to prevent extra rain being caught in very heavy rainfalls.

Rain gauges in the network are all of standard types (WMO, 1981, pp 2.6 to 2.11). The daily manual gauges are either 127-mm diameter copper gauges or 100-mm diameter plastic gauges. They are read by emptying their contents into a graduated glass cylinder, from which a reading accurate to 0.1 mm can be obtained.

Storage gauges tend to be larger versions of the daily manual gauges. Their readings tend to be less accurate because of their size and method of measurement and also because of evaporation. Evaporation is sometimes attended to by adding an oil film to a residual amount of water in the gauge. In cold places antifreeze may be used to convert snow to water. These gauges are often read by using a graduated dip stick.

Recording gauges are of two main types. The first is the tipping bucket type, usually with buckets of 0.2 mm rainfall capacity, in which tips of the bucket are timed and counted. Digital rainfall recorders generally use this method. The second type is the tilting siphon gauge. In this gauge the rainfall enters a float chamber, and as the float rises a continuous (often chart) record is gained of the rainfall. When the chamber is full a fast-action siphon empties it and the process is repeated. Most digital recording gauges are of this type.

Usually a manual gauge is placed alongside the recording gauge. The rainfall in various intervals on the chart are scaled to match this 'check-gauge' total correct any cumulative error and furnish an accurate total for the whole period.

Evaporation Measurements

Evaporation is measured by pan evaporimeters. In New Zealand the standard U.S. Class "A" (raised pan) has been adopted, but some earlier records are from other types of pan evaporimeter. The class "A" is a stainless steel pan, 254 mm deep and 1207 mm in diameter, mounted just above the ground in an open, exposed site. Temperature and wind recordings are always made near the evaporimeter, so that the evaporimeter readings can be used to estimate open-water evaporation and evapotranspiration.

The evaporation is measured by recording the water level in the evaporimeter from day to day. The difference of two successive measurements gives the pan evaporation to an accuracy of 0.1mm. The water in the evaporimeter is kept between set limits by topping-up the level as required.

Open water, or lake evaporation is often estimated as being 65 to 75% of raised pan evaporation. A value of 70% (Finkelstein, 1973) has been used in New Zealand. Such an estimation procedure may introduce errors of 10 to 20%.

Evapotranspiration is calculated using the Thornthwaite, Penman or Priestley-Taylor methods (Chapter 8).

Other Instrumentation

Temperatures and humidity, wind velocity and radiation are measured using the standard instrumentation and practice outlined in the WMO Guide to Hydrological Practices (1981) and the Guide to Climatological Practices (1983).

Climatological Data Archives

To meet the needs of a wide range of users, a new climate database, known as CLIDB, was established in New Zealand in 1992 by the National Institute of Water & Atmospheric Research.

Data Flows

Data comes into CLIDB in two forms. The first is written records from the daily rainfall stations and from climate stations that are not part of the daily climate (DLYCLI) station reporting system. The second is everything that is sent in electronic form to Wellington in real-time for forecasting, including the daily climate (DLYCLI) reports. This data flow is over five times greater in volume than the paper data stream. At present, many observations are manually recorded and then entered, in coded form, at a WOTS keyboard at the point of observation. As automatic weather stations and other data loggers become more widespread, this manual input will progressively decrease. The paper flow data are updated once a month, and the electronic data are updated daily.

Editing

Editing checks are carried out to check internal consistency in the reports, to check that the observations lie within reasonable limits, and to do some inter-station comparisons of data.

The consistency checks confirm, for example, that the minimum temperature is less than the maximum temperature. The limits checks compare data against prescribed limits for a given site and so highlight any grossly incorrect data. Inter-station comparisons are restricted to the daily rainfall network and largely reveal values entered against the wrong days.

Nevertheless, the editing in the CLIDB system does not guarantee error-free data, and the analyst should ensure that data quality is sufficient for the intended purpose. Often, a detailed examination of the output of an analysis will lead to a re-examination of some aspect of the input data.

Archiving

CLIDB consists of about eighty tables in a relational database ORACLE. About twenty tables contain the commonly used data types, and may be very large. Each of the tables normally contains one climatic element. The rainfall table, for example, has nearly 30 million rows. Indexes are extensively used in the tables to make retrievals faster.

A number of tables in CLIDB are of particular importance to users. These include tables of station histories, which describes changes in site and instrument exposures, and histories of the instruments used to gather the data. There are also a number of code tables, since some weather observations are reported in code, and statistics tables which include period-averaged data and rainfall intensity information.

The database uses standardised units. Times are all stored in UTC (Universal Time, or Greenwich Mean Time) and data can be retrieved for any period and in any time units required. Metric units are used throughout. Rainfall and evaporation are in millimetres, wind speeds are in meters per second, and temperatures are in degrees Celsius.

The database is currently about 10 gigabyte in size, and is growing at 300 megabytes per year. It does not, at 1992, contain any weather surveillance radar data, or satellite data.

In addition to CLIDB, the New Zealand Climate Databank contains a large number of chart records. These are largely weather maps and charts that give continuous records of rainfall, temperature, wind, sunshine, earth temperatures, humidity, atmospheric pressure and other items.

Data Retrieval

The main form of data retrieval from CLIDB is by using ORACLE's structured query language SQL*PLUS, which allows great flexibility in the query process. To use it, users need to know the table and table field names. ORACLE's SQL*REPORTWRITER allows users to produce well laid out reports. A catalogue table within CLIDB allows users to find what data is available. Progressively, menu driven applications will be developed for the most frequent queries.

Data from CLIDB are available at the cost of extraction and delivery to the customer. In exceptional circumstances, such as when a very large amount of data is requested for commercial use, copyright, royalty or licence fee is charged. All data are subject to copyright, and normal conventions cover matters such as supply of data to third parties, acknowledgment of source, and so forth.

Requests for data should be sent to Manager, Climate Databank, National Institute of Water & Atmospheric Research, P O Box 3047, Wellington, NZ.

Engineering Applications

Engineering design of hydraulic structures and water resource projects requires estimates of the maximum, mean, and minimum quantities of water flow. Water-retaining structures must be able to cope with floods which occur over a few days, as well as perform to expectations through droughts or sequences of flow over seasons or years. Reservoir design involves calculation of the sediment loads carried by the inflowing water, patterns of sediment deposition, and an estimate of the useful life of the storage in the reservoir (Chapter 13). Application of hydrological data usually begins with a series of checks on the integrity of the data. These checks may include a visual examination of the stage hydrograph, comparison of flow gaugings with discharge rating curves, calculation of a water balance where runoff is compared to estimates of mean catchment rainfall and evapotranspiration, and mass curve comparisons with other flow or rainfall records. Mass curves are a particularly powerful method of detecting small but persistent discrepancies in data. If two records are consistently correlated, there is a linear relationship between the two cumulative records. A small change in the correlation will alter the slope of the line and it is possible to detect changes of 3-4% in this way. The checking process often forms part of a routine quality assurance programme carried out by data collection agencies.

Discharge Computation

Estimates of mean flow are often required for locations where no flow records exist. There are a number of ways in which mean flows can be estimated for ungauged catchments:

a) from catchment mean annual rainfalls (calculated using Thiessen polygons or isohyets) and estimates of evapotranspiration (see Chapter 8), using

$$31.56Q/A = (P-E)/1000$$

where Q is the mean flow in m^3/s, A the catchment area in km^2, P the annual precipitation in mm, and E the annual evapotranspiration in mm.

b) by transposition of flow records from adjacent or upstream or downstream gauging stations, using

$$Q/Q_t = A/A_t$$

where Q_t, A_t are the mean flow and catchment area of the gauging station which is to be transposed. This is the same as applying the specific discharge of one catchment to another catchment. If strong rainfall gradients exist the estimate can be weighted by the catchment mean rainfalls (P), or alternatively by mean rainfall less evapotranspiration

$$Q/Q_t = A/A_t * P/P_t$$

c) from contour maps of runoff or specific discharge drawn for a larger region or regression equations developed for a region.

Frequency Studies

Because hydrological design standards are often based on frequency, frequency analysis is a basic tool of hydrologic design. The methods used for frequency analysis of both rainfall and discharge are similar and are described in Chapters 4 and 6 respectively. The method most commonly employed is to fit a straight line to the annual maximum values plotted on Gumbel (EV1) probability paper. The value of rainfall or discharge for the required design frequency is then read from the graph. These methods have been applied to much of the hydrological data in New Zealand and analyses of discharge are available in McKerchar and Pearson (1989) and of rainfall in Tomlinson (1980) and Coulter and Hessel (1980).

Sometimes design flows are needed for particular times of year. For example, a coffer dam may be constructed to protect some river works which are to be built over three months in the winter. What is an appropriate design flood? A related problem is the variation of flood risk throughout the year. Both can be estimated from a frequency analysis of the long-term flow record month by month.

Usually temporary works are designed for a flood with a 1 in 15 year annual exceedance probability (AEP). The combined risk of the flow being exceeded in each month must equal the probability of a 1 in 15 AEP. Thus, the required AEP (T) for each month of the year can be calculated:

probability of flow not occurring in 1 month $= (T-1)/T$

probability of flow not occurring in 3 months $= ((T-1)/T)^3$

Hence, for a three month period the probability that flow will occur $= 1-((T-1)/T)^3$
$= 1/15$

and solving for T, the AEP for each month is
$T = 44$ years

The required design flood is the 1 in 44 AEP flood calculated from the monthly maxima for the three months during which the temporary works will operate.

Similarly, the variation in flood risk throughout the year can be determined:

probability of flow not occurring in 12 months $= ((T-1)/T)^{12}$

Hence,
probability that flow will occur in 12 months $= 1-((T-1)/T)^{12}$
$= 1/15$

and solving for T, the AEP for each month is
$T = 174.4$ years

Figure 3.12 shows an example where the 1 in 174.4 AEP floods have been calculated from the monthly maxima for each month and plotted along with the 1 in 15 AEP calculated from the annual maxima.

Hydrograph Models

The basic function of a hydrograph model is to convert rainfall to runoff. Often there are more rainfall records than flow records and these data can be used to give estimates of flood frequency or to generate long-term sequences of flows for water resource studies. Because hydrograph modelling estimates flood volume over time as well as the peak discharge, it is commonly used for the design of reservoirs, flood retention structures, and other structures or schemes involving the storage of flood waters.

Methods of converting rainfall to runoff range from linear relationships based on runoff to complex models incorporating a number of physical processes (see Chapter 7).

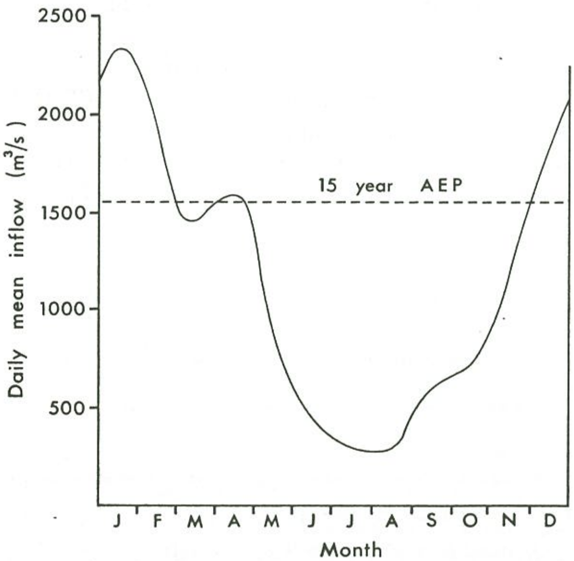


Figure 3.12 Monthly variation in flood risk for a 1 in 15 AEP flood in Lake Pukaki.

Unit hydrograph

The concept of the unit hydrograph was introduced in the 1930s and has probably been applied to more catchments than any other method of predicting flood behaviour. The definition and basic concepts are discussed in Chapter 7. The unit hydrograph describes the response of a catchment to a unit of rainfall excess spread over a unit of time. The distribution and intensity of rainfall over a catchment can vary, and this often makes the derivation of unit hydrographs difficult. However, the response of large catchments to rainfall is often damped by integrating and compensating factors, so that unit hydrograph analyses of large catchments (500 km^2) are often easier than for small. One important aspect in the application of unit hydrographs is the relationship between rainfall and rainfall excess or quickflow. At the beginning of rainfall, a portion of the rainfall is soaked up or intercepted by the ground and vegetation, and a portion contributes to streamflow. The portion contributing directly to streamflow is variously termed direct runoff, surface runoff, or quickflow. As the storm progresses the proportion of rainfall contributing to streamflow increases until 100% of the rainfall is contributing to runoff (Figure 3.13). Analysis of past storms and flood hydrographs provides an understanding of the response of the catchment to rainfall, and allows prediction of response to a given rainfall. The rainfall-runoff relationship can vary with the degree of saturation of the catchment or with time of year. In the absorbent pumice catchments of the central North Island only 5% to 40% of the rainfall contributes to quickflow, even in severe storms (Figure 3.14) whereas in saturated catchments, such as those draining the Southern Alps, 100% of the rainfall can contribute to quickflow (Figure 3.13). Dry Central Otago catchments show very little response to rainfall in summer, and even in winter when the ground is saturated and/or frozen only about 25% of the rainfall contributes to quickflow (Jowett and Thompson 1977).

Complex multi-catchment models of New Zealand river systems, based on unit hydrographs, have been used to design flood routing procedures in the Waikato (Jowett 1972) or to assess design floods for the upper Clutha catchment (Jowett and Thompson 1977). The Clutha model predicted peak lake inflows, tributary flows, and flow at Roxburgh (catchment area 15857 km^2) to $\pm 10\%$ for the majority of catchments (Jowett 1979). Multi-

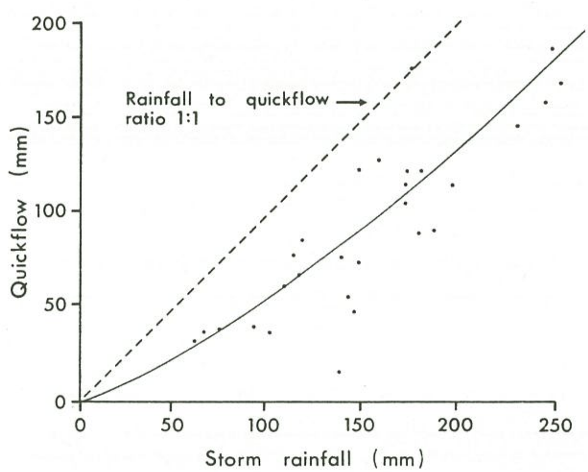


Figure 3.13 Relationship between storm rainfall and quickflow for Lake Wanaka inflow, showing how the rainfall:runoff ratio is 1:1 for high rainfalls.

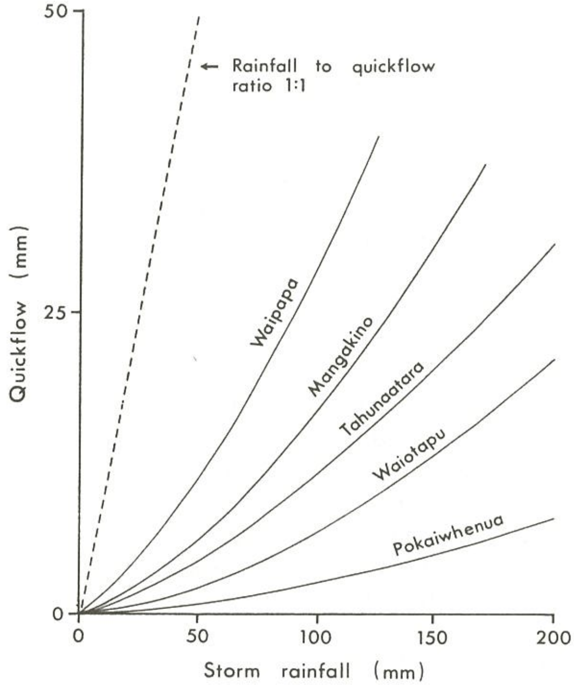


Figure 3.14 Comparison of relationships between the volume of quickflow and storm rainfalls in 6 absorbent catchments between Lake Taupo and Hamilton.

catchment unit hydrograph models allow for variation in rainfall distribution and rainfall-runoff characteristics in much the same way as distributed catchment models (see Chapter 7).

Runoff-routing models

A number of models are based on runoff-routing principles. In these models rainfall excess is routed through a series of linear reservoirs (e.g., the Nash model) or non-linear reservoirs (the Laurenson model) to the catchment outlet. The Laurenson model has been developed into a runoff-routing computer programme RORB (see Chapter 7) which is used in Australia (Institution of Engineers 1987). In this method the catchment can be subdivided to incorporate non-uniform rainfall and losses over the catchment, but the programme usually requires more detailed rainfall data than unit hydrograph models. It can also take into account existing reservoirs or natural storages and simulate the effect of planned reservoirs.

Application of RORB to the Waikanae, Otaki, Hutt, and Waimakariri Rivers, where catchment areas vary from 124 km^2 to 3210 km^2 indicates that flood peaks can be predicted with a standard error of about 14% (Griffiths et al. 1989; Pearson 1990; McKerchar 1991). In all applications the basin non-linearity constant m was assumed to be 0.8 and average values of k_c, the storage delay time for the whole basin, varied with catchment area (A), according to:

$$k_c = 1.44\ A^{0.47} \quad r^2 = 0.985.$$

This is consistent with the Australian recommendation (Laurenson and Mein 1985) of a constant of 2.2 and exponent of 0.5.

Continuous simulation models

Many models which describe mathematically the movement of water through the soils and rivers have been developed over the past 20 years. Each requires a slightly different set of parameters and input data to describe hydrological processes such as evaporation, infiltration, soil moisture storage, groundwater storage, and channel flow. Models of this form can generate continuous flow records from rainfall. Examples of continuous simulation models include the Stanford watershed model and the SHE and HYCEMOS models (Chapter 7). These models can be used to predict flood discharges as well as continuous flow hydrographs. HYCEMOS was used to assess the impact of increased rainfall from climate change on the Hutt catchment (area 427 km^2). The catchment was divided into 12 hillslopes connected by 11 gutters and 14 reaches. For each hillslope, three parameters were taken from maps and 8 soil and slope properties estimated. Channel widths, lengths, slopes, and friction constants were measured or estimated for the 25 gutters and reaches. The model was calibrated on one flood and validated on 8 other floods. Using estimates of initial soil conditions and hourly rainfall data from up to 13 automatic raingauges in or near the catchment, flood peaks were predicted with a standard error of 9.5% (Leong et al. 1992).

Flood Control and Routing

As water flows downstream discharge is altered by the storage and release of water. Reservoirs and lakes are obvious water storages, but natural channels and flood plains also retain water, especially during floods. A flood is modified in two ways as water flows downstream through storages. Firstly, there is a time lag related to the average velocity of the flow. Secondly, the flood discharge is altered or attenuated - the peak discharge occurs later downstream and the peak discharge often diminishes. The degree of attenuation depends on the amount and characteristics of storage provided by the natural channels or reservoirs.

Rivers

Most New Zealand rivers are steep compared to many rivers around the world. For steep channels, hydraulic theory shows that floods will propagate downstream with very little attenuation. This allows the use of relatively simple flood routing procedures. For example, Jowett and Thompson (1977) in routing floods from the outlets of the Clutha lakes to Lake Roxburgh assumed that there was no modification other than a time lag equivalent to a kinematic wave velocity of 2 m/s.

More recently, linear systems methods (similar to instantaneous unit hydrograph theory) using fast Fourier transforms have been used to predict downstream flood behaviour from upstream hydrographs. In the Grey catchment, this method predicted flood peaks within 12%, even though the upstream hydrographs represented only about 57% of the total flow (Goring 1984). These models have been developed and installed for flood forecasting in the upper Clutha, Ruamahanga, and Kerikeri Rivers.

An advantage of the Goring method over conventional hydrologic flood routing techniques, such as the Muskingum model (O'Donnell 1985), is that the upstream and downstream flows need not necessarily be equal in volume. A three parameter Muskingum model, which allows a laterally distributed inflow, was compared to the Goring model in the Grey River and found to give similar results (O'Donnell 1986, O'Donnell et al. 1988).

Kinematic wave theory, probably the simplest method of hydraulic flood routing (Goring 1984), has been applied to New Zealand rivers to a limited extent. In the relatively steep Waimakariri River, the kinematic model, incorporating an allowance for breaking waves (kinematic shock), was able to explain hydraulically the steepening rate of rise of flood waters between the Waimakariri Gorge bridge and the State Highway 1 bridge 51 km downstream.

Reservoirs

Reservoirs can reduce flood peaks by storing inflowing water and releasing it over a longer time; in fact some reservoirs are constructed solely for flood protection.

Flood routing through reservoirs is based on the continuity equation, i.e. what goes out equals what goes in, less the amount stored:

$$O = I - dS/dt$$

where O is the rate of outflow, I the rate of inflow, and dS/dt the rate of change in storage with time.

It is normally assumed that the water surface in a reservoir is level, and that the change in storage can be represented by the change in level times the lake area, i.e.:

$$\Delta S = \Delta L \times (A_1 + A_2)/2$$

where ΔS is the change in volume for a level change of ΔL and A_1 and A_2 are the lake areas at lake levels L and $L + \Delta L$. Many reservoirs are sufficiently steep-sided for the lake area to be considered constant. Provided the time interval is small enough, the change in lake level can be calculated from:

$$\Delta L = (I - O)\,\Delta t / (A \times 1000)$$

where L is in mm, I and O in m^3/s, the time interval Δt in secs, and A in km^2. In practice, the time interval depends on the size of lake and commonly ranges from 1 to 3 hours. Lake outflow is usually a function of lake level, for both natural lakes and reservoirs with control gates. For natural lakes, the level-discharge curve at the outlet is used to determine the outflow from the lake, whereas outflow from a reservoir with control gates is determined by gate operation and discharge based on reservoir level.

Design Flood Standards

Engineers designing hydraulic structures, from roof drainage systems to hydroelectric dams, must select a design rainfall or flow, and then design the structure so that it is capable of handling that volume of water. The design standard, often the annual exceedance probability (AEP) (see Chapter 6), is usually related to the expected life of the structure and the consequences or cost of failure. Most organisations responsible for roading, urban drainage, flood protection, and the operation of large dams have developed their own design standards. In 1992 the Hydrology Centre, DSIR (now the National Institute of Water & Atmospheric Research Ltd) carried out a survey of design standards used by New Zealand regional authorities which showed standards of protection increasing from a 1 in 5 AEP for residential property to a 1 in 100 AEP for bridges (Figure 3.15). However, other factors can influence the selection of design floods. Cost-benefit studies which compare the cost of increasing protection standards to the "benefits" or damage can assist in the selection of the "best" design value. The severity of a recent flood, the largest "historical" flood, or the acceptability of failure when loss of life is probable may also influence or set design standards. Typically, small drainage systems might be designed for events with a 1 in 5 to 20 AEP, major culverts and bridges for 1 in 50 to 100 AEP, and structures of major economic importance for 1 in 500 to 1000 AEP. Structures where failure would result in loss of life are usually designed for the probable maximum flood (PMF) - an estimate of the "highest reasonably possible flood".

Design Floods

The design process involves more than just the derivation of one design flood with an acceptably low risk of exceedance. Structures must be able to perform across a range of floods, from frequent to rare. Additional information and hydraulic skills are often required to route the flood through river channels or reservoirs and to determine the required spillway capacity, bridge opening, floodplain inundation, or amount of freeboard on stopbanks or reservoirs.

Design floods can be estimated in a variety of ways depending on the availability of hydrometric data and importance of the structure. The effort that is put into deriving design floods usually increases with the value of the structure or the damage which would result from its failure. More than one method should be used for important structures, to give both an independent check and some idea of the uncertainty of the flood estimates. Methods can also be combined. For example, a flood hydrograph for a design flood estimated by frequency analysis can be derived by scaling the output hydrograph of a runoff-routing model so that its peak equals the peak discharge derived by frequency analysis.

Methods for estimating design floods commonly employed in New Zealand are:

1. rational method and TM 61 for flood peaks for small catchments
2. regional flood frequency for flood peaks of up to 1 in 100 AEP in small to medium-sized catchments
3. rainfall-runoff models for flood peaks and hydrographs for up to 1 in 1000 AEP
4. probable maximum floods for flood peaks and hydrographs for designing structures where failure is unacceptable.

Rational method

One of the early methods of estimating design flood peaks is the rational method. If rain falls on a catchment with constant intensity, flows increase until water from the farthest point in the catchment reaches the measuring point. At this stage, the rate at which rain is falling is in equilibrium with the rate of discharge. Thus the peak flow is given by:

$$Q_p = 0.278\, C\, i\, A$$

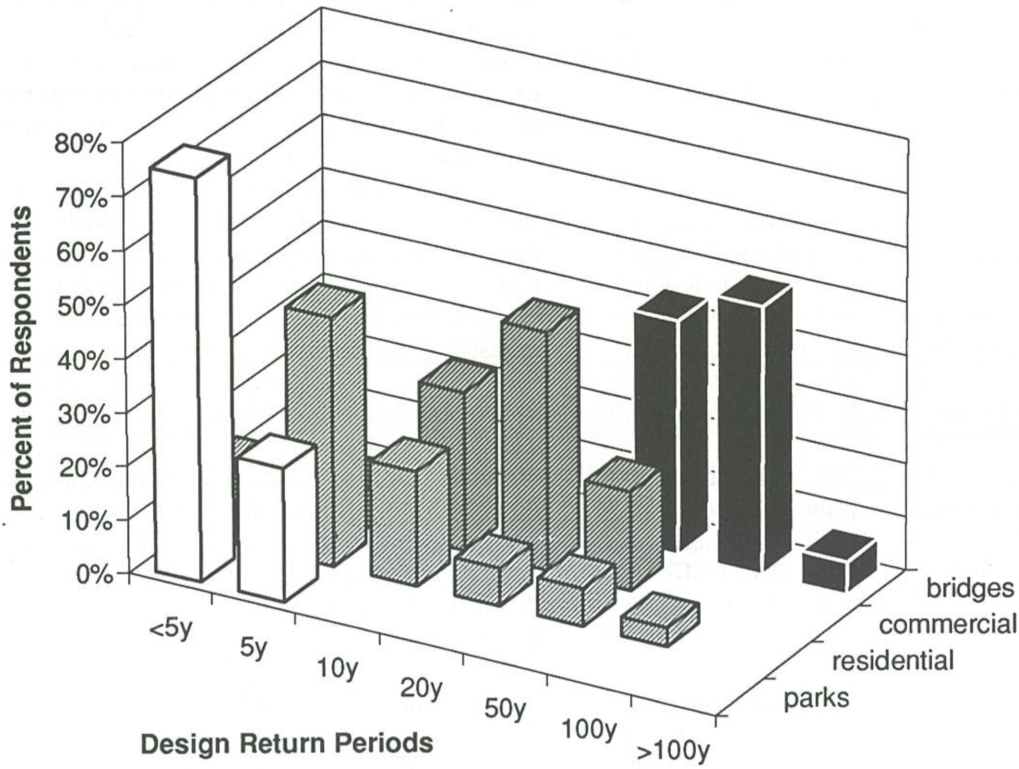


Figure 3.15 Design flood standards used by New Zealand regional authorities.

where Q_p is the peak flow (m^3/s), C is the rational coefficient, A the catchment area (km^2), and i the rainfall intensity (mm/h) for the time T_c. T_c is the time of concentration, the time required for water to flow from the farthest part of the catchment. The term "rational" arose from the use of imperial units, where 1 cubic foot per second discharge is very nearly equivalent to runoff at the rate of 1 inch per hour from 1 acre, and no constant is required in the formula.

Values of C used in New Zealand vary from 0.20 for permeable rural catchments to 0.4-0.5 for residential areas and 0.95 for roofing or impervious urban areas (R. Henderson pers. comm.). The rational method can provide preliminary estimates for larger catchments and is commonly used in flood estimation for small land and urban drainage schemes. Some New Zealand authorities have a mandatory requirement for the application of this method.

TM 61

PWD Technical Memorandum No. 61, or as it became known, TM 61 (NWASCO 1975) describes an empirical method of estimating design peak discharges for small ungauged catchments. The peak discharge is derived from the design rainfall (R) for a duration equal to the time of concentration as in the rational method. This is then multiplied by coefficients based on the shape of the catchment (S) and soil type, surface cover, channel length, and slope (C). The formula is:

$$Q_p = 0.0139\ CRSA^{3/4}$$

where A is the catchment area in km^2.

Like the rational method, the application of TM 61 is a mandatory requirement by some authorities.

Regional flood estimation

Regional methods for design flood estimation have been developed from observed flood data in a number of countries, including New Zealand, and have often been incorporated into nationally accepted guides of practice (e.g. "Australian Rainfall and Runoff" (Institution of Engineers Australia 1987) and the UK Flood Studies Report (NERC 1975)). In New Zealand, Beable and McKerchar (1982), following procedures used in the UK Flood Studies Report, developed regional frequency relationships for the ratio of flood magnitude to mean annual flood, and regression equations for the estimation of mean annual flood in ungauged catchments. This work was revised by McKerchar and Pearson (1989) using additioanl data. Regional boundaries were replaced by sets of contours covering all of New Zealand. One set allows the derivation of mean annual flood and the other gives the ratio of 1 in 100 AEP flood to mean annual flood.

This method is easy to use and is probably most suitable for catchments of 20 to 1000 km^2 for return periods of 2 to 100 years.

Rainfall-runoff methods

Unit hydrographs and runoff-routing models all convert rainfall or rainfall excess to discharge hydrographs. The methods and assumptions are more complicated than for peak discharge methods. There are a number of steps in the process of converting a rainfall of the required design frequency to a discharge hydrograph.

1. Catchment design rainfall (both temporal distribution and total) for a range of durations must be estimated from existing publications (e.g. Tomlinson 1980; Coulter and Hessel 1980) or from analysis of rainfall data.
2. Rainfall excess must be estimated by applying average rainfall-runoff relationships (e.g. Figure 3.13) or initial losses and average loss rates.
3. Rainfall excess is converted to discharge for a range of durations; the highest peak discharge is taken as the design flood; the duration of rainfall producing the highest discharge is the design flood duration.

When converting rainfall for a design AEP to a discharge of the same AEP, the temporal and spatial characteristics of the rainfall and catchment response should be "averaged" to avoid compounding the frequency of the event. The combination of unlikely or unusual occurrences can result in over-estimation of the design flood or a "conservative" estimate. Typical "conservative" procedures which compound flood frequencies are:

1. application of point rainfalls to the whole catchment with no areal reduction
2. adoption of worst temporal or areal distributions of rainfall
3. application of rainfall-runoff relationships which envelop observed values
4. an assumption of 100% rainfall excess
5. application of minimum observed loss rates and zero initial loss.

Probable maximum flood

The probable maximum flood (PMF) and the rainfall which produces it, the probable maximum precipitation (PMP), are considered to be the largest possible for the catchment. The methods of deriving PMP and PMF are similar to those used in estimating design floods by runoff-routing models. Throughout each stage of the flood estimation process, the worst possible conditions theoretically capable of occurring for the particular storm type are assumed to apply.

The process of PMF estimation begins with the determination of the PMP for a range of durations. Severe storms over the catchment and in meteorologically similar areas are examined to determine the storm type likely to produce the highest rainfall for the given duration. The storm values are then "maximised" to a uniformly high dewpoint temperature and applied to the catchment. The maximisation assumption is that the same storm could have occurred with a high dewpoint temperature (and moisture content) and that the amount of rainfall would increase accordingly. Similarly, the assumption in applying storm data from one location to another is that the storm could have occurred at the other location. The worst possible spatial and temporal variation is selected on the basis of past storms, and if values from another area are used, orographic effects must be considered. Generalised estimates of probable maximum precipitation have been produced for the United States, Australia, and New Zealand (Tomlinson and Thompson 1992). These estimates tend to be higher than estimates of PMP made for specific projects because a wider range of storms are studied with a more liberal approach to the application of storm values from other areas.

The set of PMP design storms are then converted to discharge using runoff-routing models. The worst possible antecedent conditions are assumed - usually that a large storm occurred just prior to the PMP. Past storm records give some idea of the minimum time interval between storms. Similarly, worst possible assumptions are made in flood routing. Outlet valves are assumed blocked by storm debris, power station generation is impossible because transmission lines are down, communication and power supplies are cut, and generally the flood and its consequences are worse than any other flood in living memory.

Simulation

Hydrological simulation involves mathematically simulating the operation of a water resources project in order to determine either a "best set" of operating rules or the effectiveness of a proposed design or set of rules. The operation of a simple river diversion can be simulated to estimate both the amount of water diverted and the flow hydrograph of the water remaining in the river. Simulations of irrigation schemes can use rainfall and soil moisture to determine optimum application rates. Simulated flood operating procedures for reservoirs can be used to minimize the reservoir discharge and rates of change of discharge, while maintaining reservoirs levels within design limits. Historical flow records can be used to simulate the operation of a water storage reservoir to determine the likelihood of supplying a given demand, the range of reservoir levels, and the amount and frequency of spillway discharge.

The simulation approach can describe the whole or a large part of a system. This is particularly useful for multi-use studies and for assessing complex inter-relationships. For example, hydroelectric generation in the Waitaki catchment influences daily flows, seasonal flows, and the magnitude of floods. Three large lakes regulate the flow of water, two of which, Pukaki and Tekapo, provide about 60% of New Zealand's hydroelectric storage. Peak inflow into these catchments occurs in spring and summer when heavy rain falls on the Southern Alps and accumulated snow melts. Minimum flows occur in winter when most of the precipitation falls as snow.

To meet peak electricity demands in winter, spring and summer inflows are stored in the hydroelectric reservoirs to be released gradually during the winter. The amount of water stored at any time can be compared with expected electricity demands and the historical pattern of inflows to determine the risk that the storage will not be full by the beginning of winter. If the storage is too high, water may have to be spilt to avoid overtopping the dams -this increases flood risks downstream, and is a lost opportunity to generate electricity. If the storage is too low, there may be insufficient water to meet the coming winter demand.

Alternative strategies which balance these two risks have been compared by simulation. Input data were the historical flow records for the 3 lakes and one major tributary, the design capacities of the structures, and a set of guideline levels for the storage lakes based on average electrical demand. This simulation model was used to investigate power generation and utilisation under alternative strategies and the effect on lake levels and discharges throughout the entire system. It showed, for instance that hydroelectric operation reduced the

1 in 100 AEP flood by 20% (Figure 3.16), which agreed very closely with the analysis of data for the 25 years before hydroelectric control and for the 25 years after (Jowett 1980). The same simulation model, with design flood inflows as input data rather than historical flows, was used to simulate the passage of a 1 in 1000 AEP through the system. This ensured that the flood routing procedures for each structure in the system maintained reservoir levels and spillway discharges within their design capacities.

Simulations can be used to determine "natural" flows after diversions or reservoir operations have modified the flow regime. For instance, the Tongariro power scheme has a complex flow monitoring network which makes it possible to calculate the natural flows in the Tongariro and Waikato Rivers and levels in Lake Taupo as if there were no power development.

Many of the methods described here are currently employed for design or forecasting. Methods are being continually improved or adopted from overseas, but conventional methods will continue to be used and will provide a useful baseline against which to compare new methods.

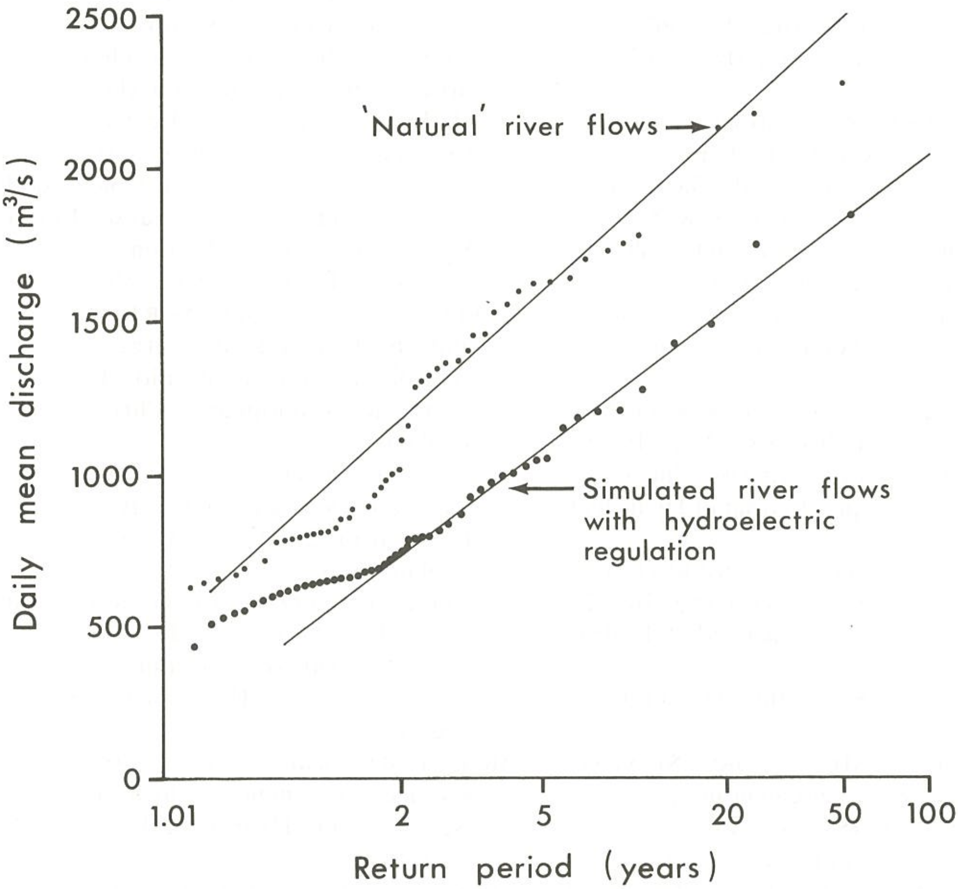


Figure 3.16 Comparison of Waitaki River "natural" floods and simulated flood discharges resulting from hydroelectric operation.

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4
Precipitation and the Atmosphere

Alaric I Tomlinson

Introduction - New Zealand's position in the global atmosphere

The atmosphere that surrounds the earth is very thin. Most weather-producing systems extend to a height of less than 14 kilometres; above this height the moisture content of the air is negligible. This thin layer of atmosphere moves, on the average, in twelve major weather systems, which form the broad outline of the atmosphere's general circulation (Palmen, 1951). From the equator to latitudes 30 degrees north and south are the north-east trade winds in the northern hemisphere and the south-east trade winds in the southern hemisphere. Just to poleward of the trade winds are belts of high pressure. Poleward again from these are zones of westerly winds which extend to latitudes of about 60 to 70 degrees. On the poleward edge of these westerlies are normally a series of depressions, and beyond those are bands of easterlies that extend around the Arctic and Antarctic high pressure areas.

For much of the year New Zealand lies in the zone of westerly winds. The main axis of this zone lies just to the south of New Zealand in an average winter, but considerably further south in summer, so that most of New Zealand lies in the high pressure belt. New Zealand's position in the general circulation is one of three major influences on its climate. The others are its location in a large area of ocean and its orography.

The axis of the high pressure belt just north of New Zealand moves during the year from about 26 degrees south in winter to about 36 degrees south in summer. The belt consists of a series of westward moving anticyclones, between which are troughs of low pressure associated with depressions in the westerly wind belt that lies to the south. This sequence of weather systems gives the constant daily change in New Zealand's weather. The interval between successive troughs of low pressure is quite variable, and only about one in two seasons experiences a fairly constant interval between troughs of low pressure. When this occurs the interval is often between five and eleven days.

Causes of Precipitation

Precipitation occurs when air rises, expands, and cools sufficiently for the water vapour in the air to reach condensation point. This is a necessary, but not quite sufficient, condition for precipitation. Three other things are also required:

1. The presence of condensation nuclei on which condensation can begin. Without these the air can become super-saturated.
2. The condensed droplets must not evaporate by passing into drier air. This effect is clearly seen when virga occurs, where rain is seen falling from a cloud base but not reaching the ground.
3. The droplets or ice crystals must grow to a size sufficient to fall to the ground.

Once very small water droplets or ice particles have formed as a result of lifting and cooling, other processes are needed to make them grow. They may grow by colliding with one another or by continued condensation. The collision process is of three types (Braham, 1959; Browning et al, 1968; Ludlam, 1980):

1. Coalescence (for liquid with liquid), which typically produces rain or drizzle.
2. Aggregation (for solid with solid), which typically produces snow.
3. Accretion (for liquid with solid), to produce ice pellets and ice grains, or hail.

Supercooled-water commonly exist in clouds when the temperature is between 0 °C and -40 °C. Because of this, in those parts of clouds that are above 0 °C only coalescence can occur, and in those parts that are below -40 °C only aggregation can occur. In the range 0 °C to -40 °C all three collision process are possible.

In clouds where supercooled water and ice crystals co-exist, one other major growth process causes ice crystals to grow at the expense of adjacent supercooled water drops. This is known as the Bergeron-Findeisen process. It is important since a large majority of precipitating clouds contain appreciable thicknesses between 0 °C and -40 °C. The process relies on the fact that the saturation vapour pressure over ice is lower than that over water.

The process is most effective in the range -10 °C to -30 °C, since above -10 °C there will generally be few ice particles and below -30 °C there will be few water droplets. If you have two adjacent particles, one ice and the other super-cooled water in air at -10 °C, then the saturation vapour pressure over the ice particle is 2.60 hPa and over the water droplet is 2.86 hPa (WMO, 1973). Neglecting the effect of the small radius of curvature of the drop, if the vapour pressure in the cloud is 2.60 hPa, the air is saturated with respect to ice but unsaturated with respect to water. Water will then evaporate from the super-cooled water drop until the air in its immediate vicinity is saturated. As this happens the air becomes super-saturated with respect to ice, and so water vapour will be deposited directly on to the ice particle. This deposition will continue until the air is saturated with respect to ice, or the water drop ceases to exist.

To produce precipitation clouds need the following characteristics:

1. If they are above 0 °C they must contain high liquid water content.
2. They must be deep and turbulent, to afford great opportunity for multiple collisions. Depth also increases the likelihood that the cloud will include an appreciable thickness in the 0 °C to -40 °C layer where the Bergeren-Findeisen process can operate.
3. They must have existed for some time, to increase the opportunities for multiple collisions.

Major Weather Systems

Major weather systems account for much of the precipitation that falls over New Zealand. They are one of three major producers of precipitation, the other two being orographic uplift and mesoscale weather systems such as thunderstorms and line squalls. Mesoscale systems are generally too small to be shown on the weather maps seen in newspapers and on TV, but can be major precipitation producers.

All three types rarely occur alone. Orographic uplift often occurs in combination with each of the other two types, and mesoscale systems are often embedded in the major weather systems.

The major weather systems are often called synoptic scale systems. Synoptic scale systems are those revealed on synoptic charts, and include depressions, frontal zones and troughs of low pressure. They vary greatly in structure, size, intensity, and the length of time they persist. They have quite short life cycles and are constantly forming, growing, decaying, and dying. As they interact with each other or with mesoscale systems they may intensify or otherwise change their characteristics.

Typically a synoptic scale system has a life span of 3 to 10 days. During this time it will produce rainfall over an area of one thousand to tens of thousands of square kilometres. This rainfall will very rarely be uniformly spread, because of the cellular nature of many precipitation producing clouds and because of interaction with mesoscale systems. This irregular distribution is particularly typical of heavy rainfall.

There are many texts that describe these systems, for example, Barry and Chorley (1982), Atkinson (1981), Sumner (1988).

Thunderstorms and Convective Precipitation

Thunderstorms are important to New Zealand precipitation in that they contribute to local heavy falls. In Otago and particularly central Otago, however, they are the major contributor to total precipitation in summer. Thunderstorms are the principal cause of flash flooding in New Zealand, as they are the major contributors to all heavy rainfalls of durations up to a few hours. Their role in generating probable maximum precipitation is described later in this chapter, and more fully in Tomlinson and Thompson (1992).

Convective precipitation comes from single or groups of towering cumulus or cumulonimbus clouds. Such clouds are cellular in nature, and a typical thunderstorm may consist of several cells that extend over an area of a few tens of square kilometres. As the thunderstorm moves the cells in it continually decay and new cells develop, and the whole storm may traverse a few hundred square kilometres. The structure and nature of these clouds is well described in Magono (1980), Newton and Frankhauser (1975) and Ludlam (1980).

Most places in New Zealand have thunderstorms on between 5 and 20 days per year (Tomlinson, 1976), with highest numbers in the mountainous areas of the West Coast. Thunderstorms tend to occur in large, unstable air masses that commonly take several days to traverse New Zealand. Because of this, when thunderstorms occur they tend to be scattered over regions, or even the whole country, and thundery conditions may continue for a few days. This is reflected in the statistics of occurrence of thunderstorms as high spatial correlations and strong temporal persistence. Temporal persistence means, essentially, that if thunder occurs at a point today it is more likely to occur at that point tomorrow than if it has not occurred today.

Thunderstorms are violent weather phenomena, and have a major impact on human activities (Kessler, 1981). They produce and are associated with floods, severe wind squalls, hail, tornadoes and lightning. All of these are moderately common in New Zealand. On average there are at least twenty tornadoes per year in New Zealand but numbers vary greatly from year to year (Tomlinson and Nicol, 1976). Like thunderstorms their occurrence shows strong temporal persistence, and indeed several tornadoes can be associated with a single thunderstorm, or thunderstorm cluster.

The occurrence and nature of damaging hail is described by Neale (1977). Hailstones the size of tennis balls have been observed on a number of occasions. The structure of hailstorms is described by Browning (1977).

Severe wind squalls are common in the vicinity of thunderstorms, and are a particular danger to aircraft operations at and near airports. NZ Meteorological Service (1982) discusses this phenomenon at Auckland Airport.

Lightning has not been well studied in New Zealand, largely because it is difficult to observe in a consistent manner. The economic and social impact of lightning has been low; one exception is its effect on electricity transmission. Lightning counters have been progressively installed on automatic weather stations around the country since 1988.

New Zealand Rainfall

Rainfall in New Zealand arises from three interacting causes. First and foremost are the weather systems discussed earlier. Second is the maritime position of New Zealand. Every airstream that approaches this country has traversed several thousand kilometres of ocean and is moisture laden. Air coming from the warmer ocean to the north may be carrying a great deal of moisture.

The third cause is orography. New Zealand is a mountainous country with one third of the land area above 1000 metres. These mountains, and their orientation to the predominant wind flows, strongly control the amount and distribution of rainfall. From the surface to around 5000 metres,

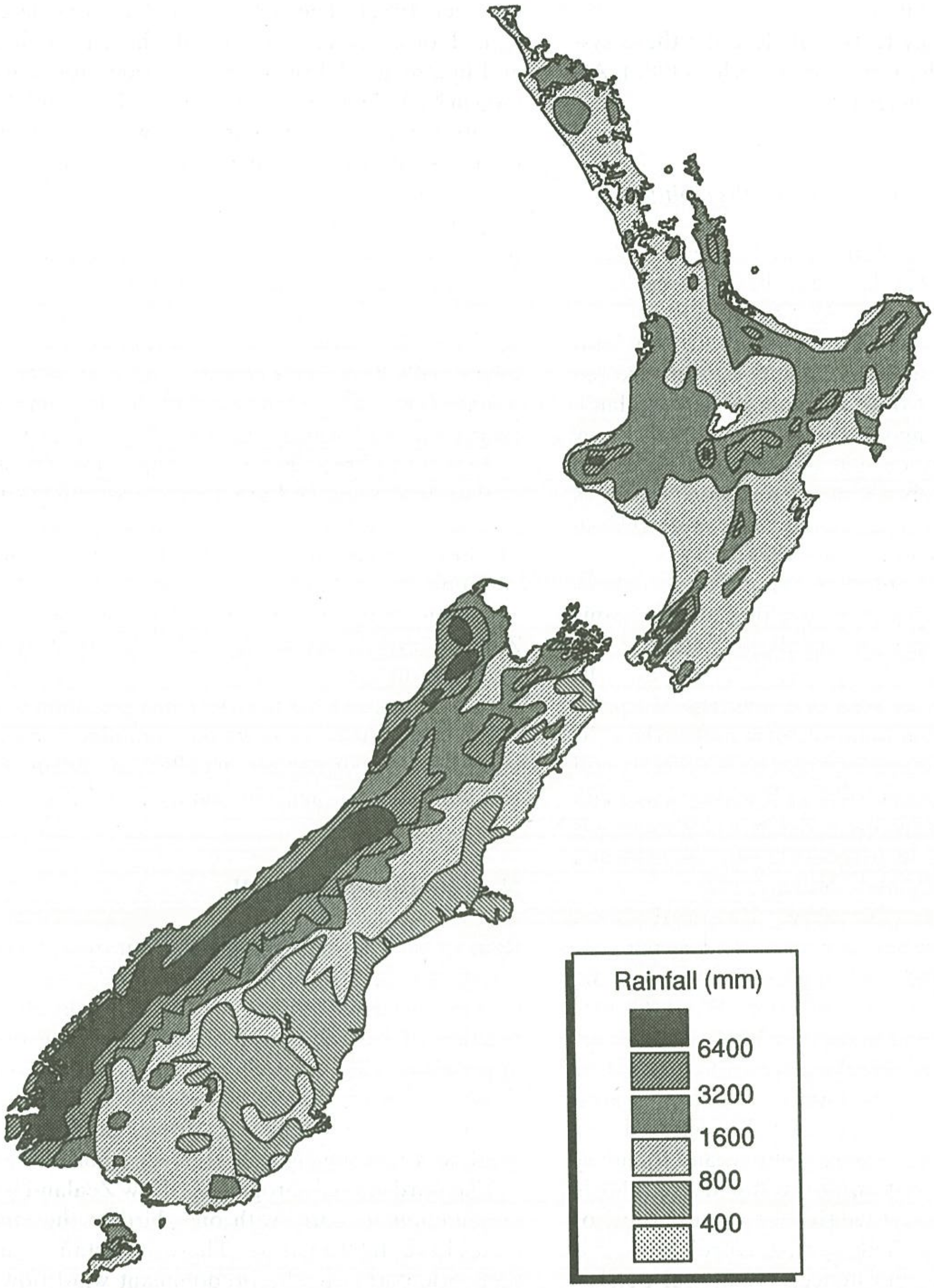


Figure 4.1 Average annual rainfall (mm) over New Zealand.

which contains most of the moisture in the atmosphere, the average wind flow over New Zealand is west-southwest. The main mountain ranges intercept this moist wind flow and the resultant orographic uplift ensures that the western part of the country is generally wetter than the eastern part, although there is substantial variability about this average trend (Figure 4.1).

Average Rainfall and its Variability

Not only do the mountains greatly augment the rainfall in some areas, but they also significantly decrease it in other areas. If New Zealand were absent, the rain over the ocean as its latitude and longitude would be 600 to 800 mm per year, compared to the actual values of about 400 to 11,000 mm. The wettest areas of the country are on the western slopes of the Southern Alps and the driest are in central Otago. Here, surrounding mountains tend to extract moisture from the winds blowing from any direction before they reach the area.

The rainfall maximum on the western slopes of the main mountain ranges does not occur at the top of the ranges (Griffiths and McSaveney, 1983). In the Southern Alps, for example, the maximum rainfall is at an altitude of 1200 to 1700 metres. Most of the moisture in the atmosphere is at the lowest levels, and so at altitudes above 1700 metres the moisture in the air coming from below this height has been largely rained out, and the moisture in air coming from above this height is insufficient to allow large rainfalls.

Most of the populated part of the country has annual average precipitation between 800 and 1500 mm. This rainfall is fairly evenly distributed through the year (Figure 4.2), apart from a tendency for a small winter maximum in many areas. The notable exception is a slight summer maximum in parts of south Canterbury and Otago.

Rainfall in New Zealand occurs overall on about one day in three. However, in wet seasons it may occur more frequently than one day in two, and in the wetter parts of the country it is also more frequent. The average number of days per year on which 1 mm or more of rainfall occurs (Table 4.1) ranges from less than 80 to more than 180 days.

Maps of annual rainfall, coefficient of variation of annual rainfall,, number of raindays, and rainfall reliability (expressed by percentile values) have been published by NZ Meteorological Service (1985), at a scale of 1:1,000,000.

Although New Zealand has a relatively reliable rainfall from year to year, there is still marked variability between years (Figure 4.3) and even more so between seasons. Eastern parts of the country from north Otago to Northland experience the greatest variability. The most reliable rainfall is in the west and south, where rain-producing weather systems in the prevailing westerlies form a regular weather pattern. This reliability pattern is accentuated for monthly rainfalls (Table 4.2). For example, the ninety-fifth percentile value of January rainfall is four times the median value at Hastings, but only about twice the median at Greymouth.

Duration of Rainfall

The actual duration of rainfall is quite short - it rains just 5% of the time at Nelson and 6.5% of the time at Auckland, for rainfall as defined in Table 4.3.

For places with average annual rainfall between 500 and 2,800 mm the following simple relationship provides a good approximation :

$$Y = (X/8) + 80$$

where Y is the average number of hours per year of rainfall (Table 4.3) and X is the average annual rainfall in millimetres.

Temporal Patterns of Rainfall

Although New Zealand's rainfall is on average quite reliable compared to many places in the world, it varies markedly over different periods of time. In general, short period variations are more localised than longer period variations, because they result from local phenomena. Longer period variations tend to result from very large scale phenomena, or even extra-terrestrial influences

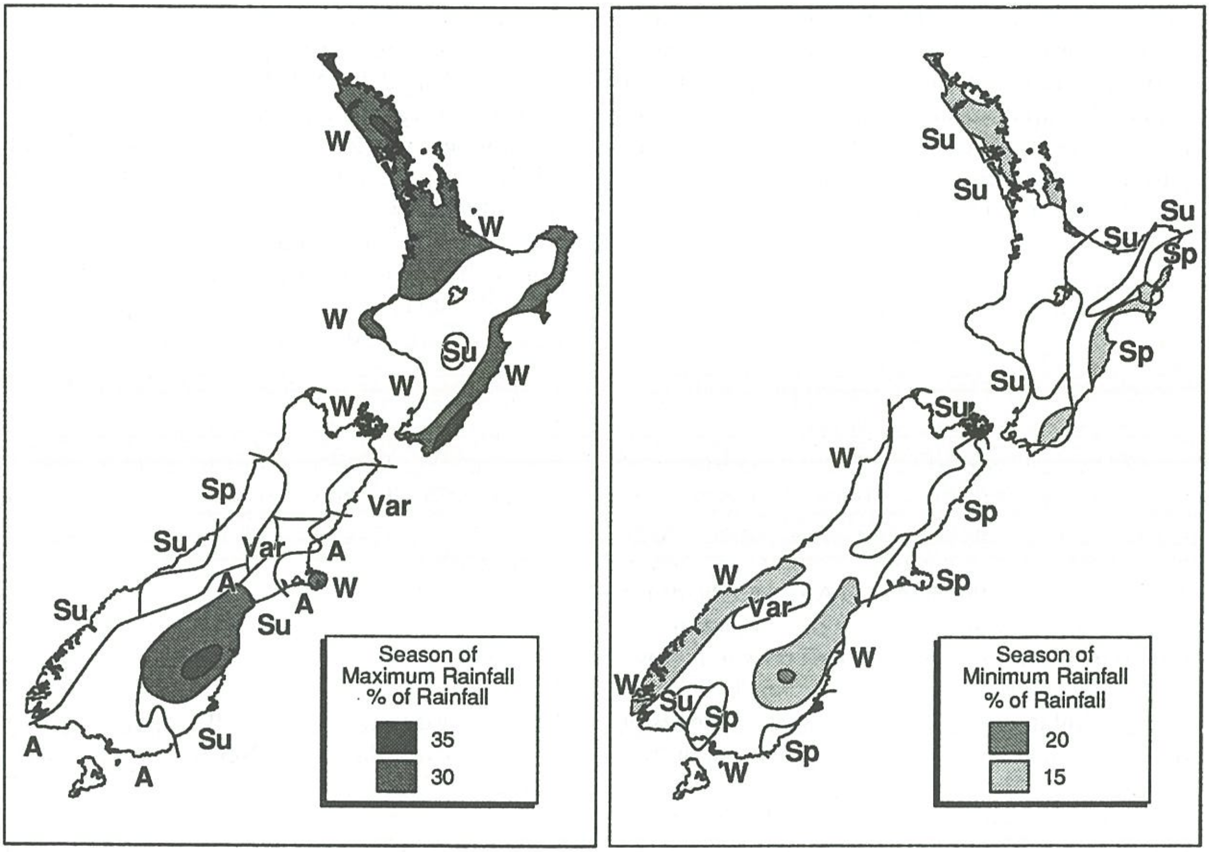


Figure 4.2 Seasons of maximum and minimum rainfall. Summer (SU): Dec/Jan/Feb; Autumn (A): Mar/Apr/May; Winter (W):June/July/Aug: Spring (Sp): Sept/Oct/Nov. Areas are shown which receive more than 30 and 35% of total rainfall in the season of maximum fall and less than 15 and 20% of total rainfall in the season of minimum fall.

such as variations in solar output.

At the smallest time scale, the common rain producing systems such as showers from cumulus clouds may cause several periods of rainfall in a day. These short cycles of rainfall show no regular pattern, and merely reflect the random scatter of shower clouds in an airstream.

At a slightly longer time scale, a few areas in New Zealand show a diurnal variation of rainfall. Where it does occur the variation is small, since rainfall tends to be evenly distributed throughout the 24 hours. However, two types of diurnal variation are detectable.

The first arises in areas where a humid afternoon sea breeze combines with afternoon heating of the land to encourage the development of large cumulus or cumulonimbus clouds. Coastal hill country may experience this effect from October to April (more commonly December to March), with the result that around 33% of rain may fall between noon and 6 pm, rather than 25%.

The second variation is of the same, or lesser, magnitude and occurs between midnight and 6 am in some coastal areas in winter. Its cause is not altogether clear, but it may result from convergence and uplift of air as katabatic winds meet onshore winds from the sea in the early morning.

Station	Annual Rainfall (mm)	Rain days (1.0 mm or more)
Kaitaia Airport	1420	138
Kerikeri	1680	135
Auckland (Albert Park)	1185	140
Tauranga Airport	1350	118
Hamilton	1200	131
Rotorua	1490	123
Taupo	1180	122
Waiouru	1050	137
Gisborne	1060	113
Napier	825	95
New Plymouth	1530	144
Wanganui	905	115
Palmerston North	995	126
Masterton	970	124
Wellington	1240	125
Nelson	985	99
Blenheim	640	81
Westport	2190	168
Hokitika	2785	168
Milford Sound	6265	182
Hanmer Springs	1165	114
Christchurch	665	87
Timaru	585	81
Queenstown	805	92
Alexandra	345	65
Dunedin	785	120
Gore	835	136
Invercargill	1035	157

Table 4.1 Average annual rainfall and average annual number of rain days.

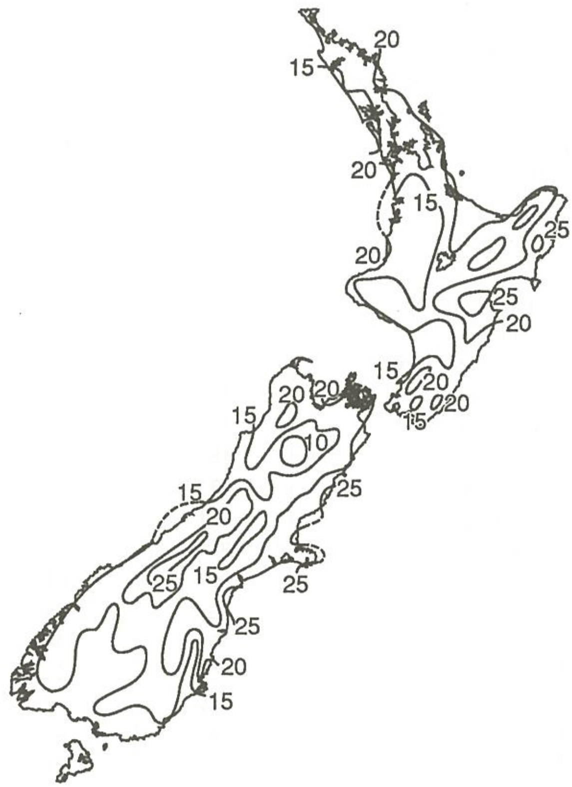


Figure 4.3 Coefficient of variation (%) of annual average rainfall. Higher values indicate greater variability.

At longer time scales again, we are all familiar with runs of wet weekends. Such weekly cycles are observed from time to time, and in exceptional cases they may persist for up to ten weeks before breaking down. The weather pattern settles into a regular sequence of anticyclones and intervening troughs of low pressure, which pass over New Zealand every seven days. The planetary atmospheric wave pattern can settle into this mode, with regular patterns of rainfalls at periods ranging from 5 to 10 days, but the periodicity is most noticeable at 7 days because it then affects the same day of the week.

At periods beyond 10 days there are a number of other cycles. They generally have varying periods and amplitudes, are erratic, and are often quite transient. Notable and sometimes detectable are:

1. A 28-29 day cycle variously ascribed to the period of rotation of the sun, which may cause variations in solar flux, or to the lunar cycle, which has tidal influences in the atmosphere.
2. A quasi-biennial cycle related to the quasi-biennial oscillation in the atmosphere, which is itself a well established phenomenon.
3. An oscillation related to the Southern Oscillation. Years at the points of maximum amplitude of this oscillation have quite identifiable rainfall characteristics over New

Station	January			July		
	95%	50%	5%	95%	50%	5%
Kaitaia Airport	291	60	15	278	145	50
Kerikeri	359	62	6	355	179	35
Auckland (Albert Park)	178	57	10	232	131	35
Tauranga Airport	225	83	9	249	122	48
Hamilton	172	67	15	213	120	50
Rotorua	257	90	11	242	128	43
Taupo	179	66	14	227	107	32
Gisborne	152	66	11	256	98	29
Hastings	198	50	5	216	83	21
New Plymouth	255	95	18	263	153	38
Wanganui	132	71	16	144	79	28
Palmerston North	167	76	26	167	80	32
Masterton	153	54	14	210	99	34
Wellington	193	72	6	256	141	56
Nelson	188	56	7	157	89	25
Blenheim	131	43	6	125	56	13
Westport	291	142	50	304	192	52
Greymouth	388	179	70	340	188	61
Milford Sound	1205	561	149	740	325	84
Hanmer Springs	193	84	27	259	94	21
Christchurch	121	47	15	169	62	13
Timaru	125	52	21	125	34	7
Queenstown	132	69	19	125	51	9
Alexandra	87	40	11	41	13	1
Dunedin	141	67	30	130	52	22
Gore	143	78	26	113	47	21
Invercargill	165	81	34	128	64	29

Table 4.2 Five, fifty and ninety-five percentile values for January and July rainfalls (mm).

Place	Hours of rainfall(Y)
Kaitaia	234
Auckland	235
Rotorua	253
Gisborne	203
New Plymouth	262
Paraparaumu	193
Wellington	256
Nelson	180
Hokitika	447
Kaikoura	166
Christchurch	145
Dunedin	173
Invercargill	253

Table 4.3 Average number of hours per year of rainfall of 0.5 mm per hour (or greater) between 7.00 am and 5.00 pm.

Zealand (Gordon, 1985). The period here is often 3 to 8 years.

4. A cycle of about ten years related to the 10-11 year sunspot cycle (Tomlinson, 1981).
5. Cycles of 18 to 22 years, ascribed to a number of causes and described by Vines (1984) and others.

For periods longer than 20 years, rainfall records in New Zealand are too short to allow analysis of variations. However, variations at longer periods do occur and can strongly affect average rainfalls. For example, average rainfalls during the periods 1921 to 1950 and 1951 to 1980 differ by 4% over large areas of New Zealand, and by up to 8% in some smaller areas.

Rainfall patterns commonly demonstrate a phenomenon called persistence (Finkelstein, 1967). Persistence relates directly to the weather patterns causing rainfall. For example a major trough of low pressure can take several days to cross the country. Each day will tend to be rainy, and so runs of wet days occur. In Wellington, for example, there are on average 125 days per year with 1 mm or more of rainfall. If such raindays were randomly distributed, the probability that any day will be such a rainday is 0.34. However, if the previous day was a rainday this probability rises to about 0.50.

The persistence effect is also observed, but weaker, in monthly rainfall figures, but is generally absent for annual rainfalls.

Storm Rainfalls

In many parts of New Zealand storm rainfalls have recognisable patterns, which are related to their causes. Salinger et al (1986) analysed the spatial distribution of storm rainfalls in the Wellington area and Trewinnard and Tomlinson (1986) carried out a similar analysis for central Canterbury.anterbury In each area five characteristic patterns of rainfall were identified, which in most years accounted for 85 to 90% of the variance in daily rainfalls. These patterns can be related to the weather systems that produce them.

Extreme Rainfalls

The effects of extreme rainfalls, including floods and landslips, can be both dangerous and destructive. Efforts must be made to forecast them, and to design structures to withstand their impact. Forecasting in New Zealand is aided by weather satellite information and weather surveillance radar information.

A national set of frequency estimates for high-intensity rainfalls is given in Tomlinson (1984). This publication enables designers to easily estimate the size of rainfalls of various recurrence intervals, for durations between 5 minutes and 72 hours. The publication gives maps at 1:1,000,000 scale, of the 5-year return period rainfalls of durations 10 minutes, 1 hour, 6 hours, and 24 hours, and a set of tables and graphs to convert these values to other durations and return periods. Lengths of records used in the analysis of high-intensity rainfalls are insufficient to estimate return periods over 100 years.

For point values of high intensity design rainfalls from frequency analysis, it has been standard practice to use overseas analyses as described in Tomlinson (1980, revised 1984) to produce areal values and to distribute the rainfalls through the duration of storms. Figure 4.4 is a set of depth-area curves derived from an analysis of the heaviest storm over an area of 1,000 square kilometres in the Wellington region over a period of 30 years. These were derived from 13 recording and 37 manual daily gauges in the area, and data over the thirty years 1950 to 1979. They allow point high intensity rainfall values to be converted to average values over catchments.

Figure 4.5 gives a temporal design pattern from the same Wellington region analysis. This figure allows the total design rainfall to be distributed through the duration of a storm. It is notable that Figures 4.4 and 4.5 show results quite similar to those reviewed in Tomlinson (1984).

In some circumstances the consequences of the failure of engineering structures such as dams or

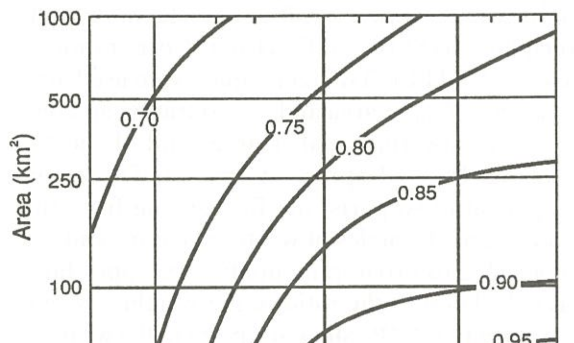


Figure 4.4 Depth-area curves of high intensity rainfalls. This shows, for example that for a 6 hour rainfall the 250km^2 value is 81% of the point value.

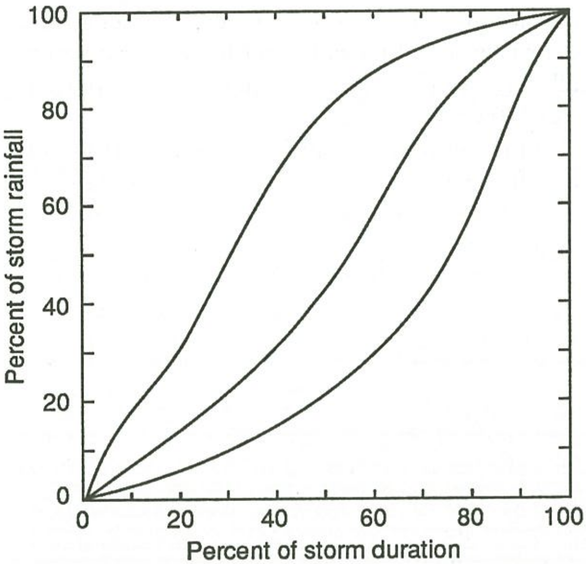


Figure 4.5 Average rainfall accumulation during 17 annual maximum storms in the Wellington area (middle curve). The upper and lower curves envelope the 17 individual storm curves.

stopbanks are so serious that information on rainfalls with longer return periods are needed. To this end Tomlinson and Thompson (1992) produced national estimates of probable maximum precipitation (PMP) for Electricity Corporation of New Zealand Ltd. Two techniques were used, one for storms of large area and long duration and one for storms covering smaller areas and of shorter duration. For widespread storms, rainfalls could be split into two parts, the first arising from the atmospheric dynamics of weather systems and the second due to orographic uplift. An upper limit was calculated to the ratio of the weather system component of PMP values at a point to the weather system component of 100-year return period rainfalls. The weather system component of PMP could then be mapped. This component was then converted to a total PMP value by adding the part due to orographic uplift. The process was completed by developing algorithms and nomograms to convert the values to different durations of rainfall and catchment areas, and to spatially and temporally distribute the estimated PMP values.

For PMP estimates for small areas and short durations, a variation was used of the thunderstorm model developed by the Australian Bureau of Meteorology (1988). For the north of the North Island this gave point estimates of PMP for one, two, and six hours of 200, 300 and 420 mm respectively. Estimates of PMP decrease southwards, to 70% of those values at Invercargill. Nomograms were devised to adjust for duration and area of rainfall, and to spatially and temporally distribute the PMP estimates.

Droughts and Dry Spells

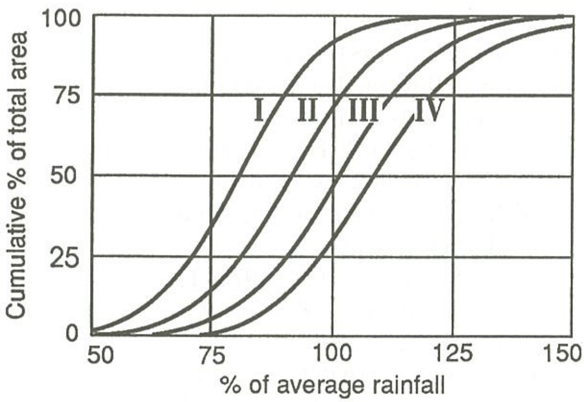
The successful analysis of droughts remains a problem, in part because of the difficulty of defining what a drought is. Their causes are not as clear as those of floods. They appear slowly, spread amorphously, and can end in various ways. We tend to know when we have one but find it difficult to know when one is beginning or ending.

Droughts arise from the persistent absence of rain-producing systems, or, perhaps more correctly, from the persistent presence of anticyclones, and in eastern areas from the persistence of westerly or northwesterly winds. These causes are large scale ones, which affect substantial areas simultaneously. Hence, whereas floods are normally localised, droughts are of regional extent.

Droughts can be viewed from several perspectives. From the meteorological viewpoint we consider rainfall and this gives the vertical perspective. From the hydrological viewpoint we consider such things as flows in waterways and this gives the horizontal perspective. From the agricultural viewpoint we consider production and this gives the producers' perspective. From the economic viewpoint we consider commerce and this gives the bottom line perspective.

With all of these views we need to consider all of the following :

- ☐ the frequency of events
- ☐ accumulated effects
- ☐ temporal distribution
- ☐ spatial impacts



Line	NZ rainfall % of normal	% of years
I	80/90	11.5
II	90/100	32.5
III	100/110	42.5
IV	110/120	12.5

(note: 1% less than 80)

Figure 4.6 Cumulative distributions of the total area of New Zealand with various percentages of rainfall. Each curve relates to a different value of the average rainfall over New Zealand.

Such a matrix of views and analyses shows that drought is a complex item whose definition is not a simple matter.

In terms of rainfall alone, a drought is a smaller magnitude effect than a flood, since rainfalls several hundred percent above average cause floods, whereas rainfall during a drought cannot be less than one hundred percent below average. Large amounts of water are stored in the ground, lakes, and rivers, and their depletion takes some considerable time. Droughts are therefore longer term phenomena than floods.

These ideas of the nature of droughts carry through into annual rainfall distributions. For example, the percentage of New Zealand that is wetter than average in a wet year is less than the percentage of New Zealand that is dry in a dry year (Figure 4.6). In years when the average rainfall over the whole country is 10 to 20% below normal, over 90% of the country tends to have below average rainfall, whereas when the average rainfall over the whole country is 10 to 20% above normal, 30% of the country has below average rainfall.

Some droughts can be adequately analysed in terms of frequency statistics of rainfall or soil moisture. Such statistics are relatively straight forward to compute for any particular place, but the task becomes more difficult if such statistics must also deal with the growing or decreasing areal extent of a drought.

Successful drought analysis remains a prime challenge for hydrologists !

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5
Snow and Glacier Hydrology

Blair Fitzharris, Ian Owens and Trevor Chinn

Role of Snow and Ice in the Hydrologic Cycle

Only about 6% of the total precipitation over the entire globe falls as snow, so at first glance snow and ice do not seem an important part of hydrology. Storage of water in solid form, however, has a powerful influence on runoff over both short and long time scales. For example, threequarters of Earth's fresh water is in the form of snow and ice, the vast bulk of which is locked up in the ice sheets of Antarctica and Greenland (Barry 1969). If melted, this ice would lift sea levels by over 80 m (Kuhn 1989). Ice cycles through these great ice sheets so slowly that the oldest ice exceeds 200,000 years in age, and extracted cores have been used to reconstruct climatic conditions back in time to beyond the last 150,000 years.

The greatest variations in flows derived from meltwater stem from the annual cycles of seasonal snow cover. In the northern hemisphere, snow cover on land shrinks from about 50 million km^2 in winter to 10 million km^2 in summer. The southern hemisphere is different; here seasonal changes are dominated by variations in sea ice cover around the Antarctic which annually oscillates from 15 to 35 million km^2 (Kukla 1978). In New Zealand snow accumulation and melt in alpine areas feeds a strong seasonal pulse to alpine stream flows. Although precipitation is almost uniformly distributed throughout the year, winter build up of snow holds back water from runoff to be released later as snow melts during spring and early summer.

The oldest ice within a glacier may be tens to hundreds of years old, depending on the size and flow rate of the glacier. New Zealand's Tasman Glacier probably contains our oldest ice, calculated to be some 500 to 800 years old. The mass of ice held in glaciers continually grows and diminishes over decades following climatic variations. When glaciers expand, as they did over the past five hundred years or so of the "Little Ice Age", which ended about 1850, they store water and diminish river flows. When they recede, as they have done this century due to global warming, water released from long term storage is added to river flows. These additions amount to less than one tenth of the water added by seasonal melt of snow and ice, but worldwide, the sum gained from dwindling glaciers does have a detectable effect on sea level.

Snow and Ice in New Zealand

Glaciers

New Zealand's Southern Alps have extensive areas of perennial snow and ice, with glaciers as large as those of Europe, and twice as many as in Switzerland. With the glaciers of the South American Andes, they are the only significant areas of temperate ice in the Southern Hemisphere (Figure 5.1). Their fluctuations have left a detailed record

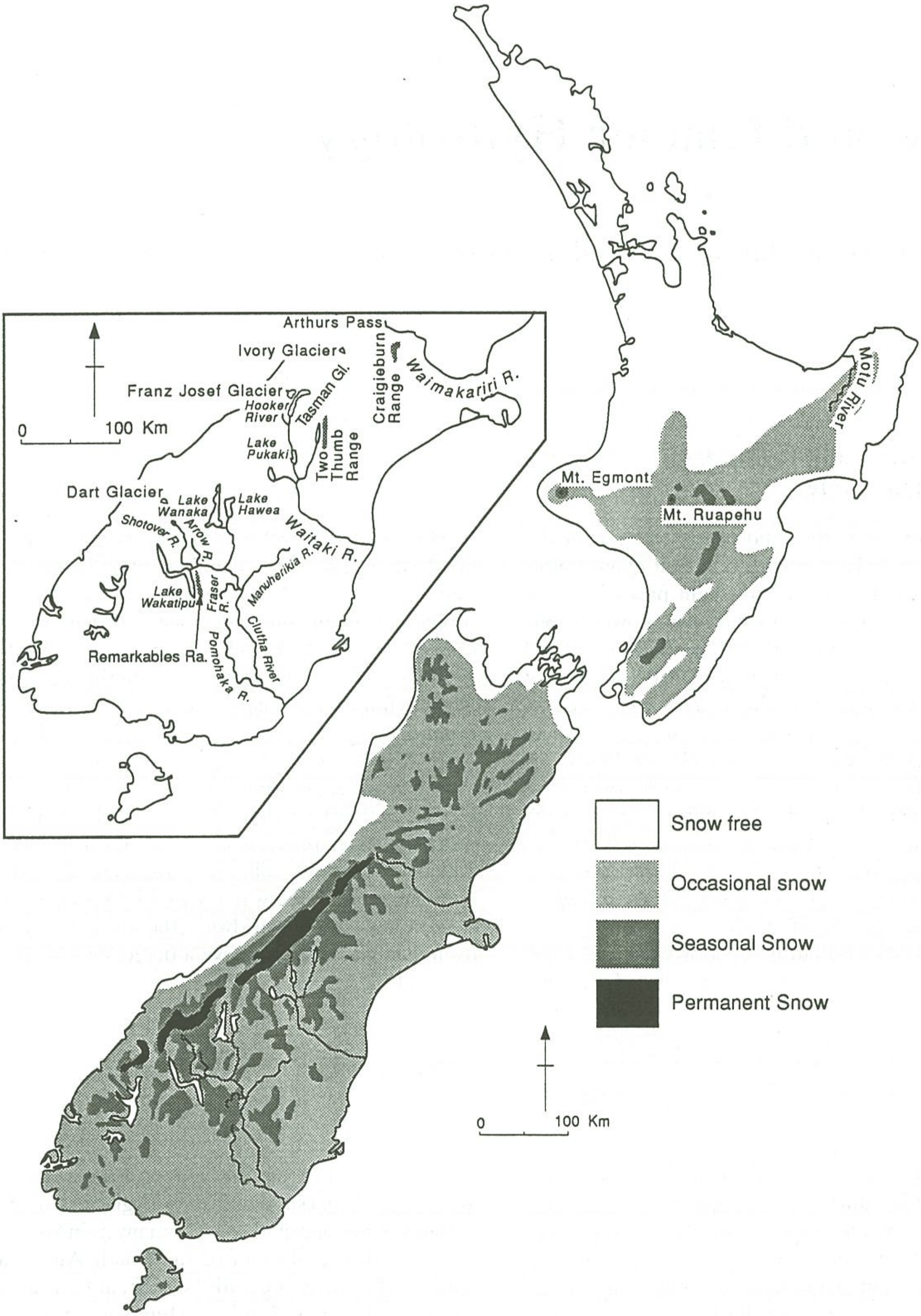

Figure 5.1 Snow and ice covered areas of New Zealand and names of locations referred to in the text (modified from N.Z. Committee for the I.H.D. 1969).

of climate change, and there is a reasonable recorded history of their behaviour since the middle of the 19th century.

Seasonal Snow

Our seasonal snow cover provides a water resource currently used mainly for hydroelectric generation and irrigation. Near the Main Divide of the Southern Alps, annual snow accumulation exceeds 4000 mm water equivalent, but it is generally less than 1000 mm in the eastern mountains. Much tourism and the ski industry is based on snow and ice. Snow is also responsible for hazards such as floods and avalanches, and its distribution has critical ecological implications for many of our national parks and natural mountain areas.

Meltwater Runoff

Each winter, rivers normally fall to their lowest levels of the year as the snowpack accumulates in the mountains (Fitzharris 1979). This stored water is later released by spring and summer thaw to markedly increase the flow of alpine rivers. For hydro-electric generation, it is therefore very desirable to control lakes, so that this enhanced spring flow can be held over summer to be released to meet winter power demands. Climate is never constant, so that in some wet years surplus water has to be spilled without being used for generating, while in dry years there is a shortage of storage in the lakes. Thus the seasonal control exerted by accumulation and melt of snow has important economic implications in this country, where hydro generation makes up 61% of total capacity and meets 77% of all electricity demand.

Climate Change

Recent concern about global warming has focussed attention on the future behaviour of snow cover and glaciers. Half of the predicted sea level rise of about 0.65 m for the next century could come from diminishing snow and ice volumes in temperate mountain areas. In this context, New Zealand's mid latitude, oceanic location puts it in an important position for study of the past and to predict future climate change from perennial snow and ice. In particular, our mid-latitude location means that the glaciers are well positioned to detect shifts in global atmospheric circulation patterns.

How Much Permanent Snow and Ice Lies in our Mountains?

Inventory of our Glaciers

A recent inventory count has tallied some 3153 glaciers in New Zealand (Chinn 1989a) which in total cover an area of 116 km^2, or about 5% of the South Island (Figure 5.1). Mount Ruapehu has the only glaciers in the North Island, with 18 small glaciers covering a total of 507 ha. South Island glaciers are scattered from southern Fiordland to the ranges about the Nelson lakes, and are fairly evenly distributed along both sides of the Main Divide, with 49% of the ice area to the east and 45% to the west (Figure 5.2a). The glaciers contain a calculated total volume of some 53.3 km^3 (Figure 5.2b), and as ice volume is related exponentionally to glacier size, the greater part is contained in the few largest glaciers, with the single largest fraction of this ice contained in the 29-km long Tasman Glacier, New Zealand's largest.

Snowlines

The snowline is the elevation of the boundary separating snow covered land from the lower uncovered land. But beyond this simple definition complications begin. The seasonal snowline at any given time is transient and fluctuates with the weather. An average winter snowline is a more stable value. For the higher mountains, there is the maximum height to which the snowline retreats at the end of each summer. This is called the regional or permanent snowline, and above this elevation the snow that lingers on into the next winter becomes part of a glacier. On a glacier at the end of

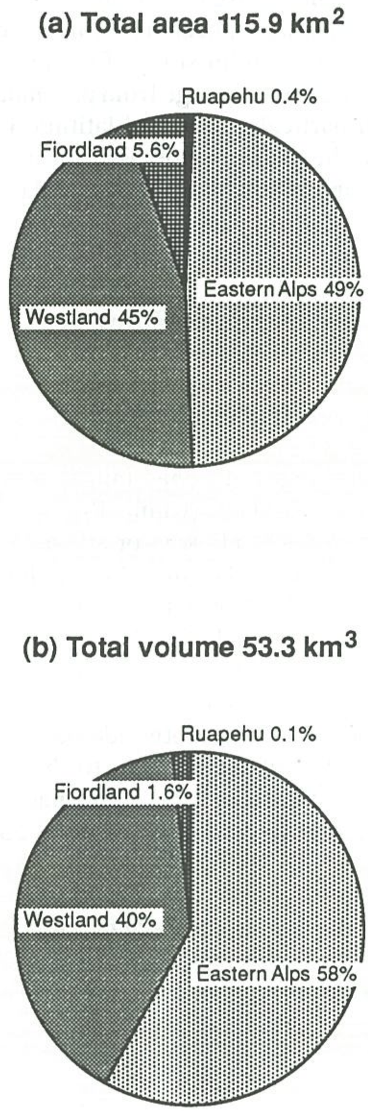

Figure 5.2 Regional distribution of the New Zealand glaciers by (a) area and (b) volume.

summer, this line is called the glacial snowline, and above this elevation is some winter snow left over from the melt season. It is this snow that feeds the glacier. Mountains below this level will not have glaciers. On a worldwide scale the snowline forms a planar surface in the atmosphere that envelopes the earth. This surface descends towards the poles from about 4500 m elevation over the equator to intersect sea level at about the Antarctic Circle. Only those mountains which rise through this surface will carry glaciers. In the North Island of New Zealand, it clips the top of Mt Ruapehu at 2440 m, passes over Arthur's Pass at 1900 m and descends to 1500 m over southern Fiordland. Its elevation is very sensitive to precipitation and it wobbles markedly over the Southern Alps. From west to east, across the Main Divide, it is distorted by the phenomenal precipitation gradient. Depressed to 1500 m under the 15 m per year precipitation totals of Westland, the snowline rises steeply upward to the east at up to 40 m/km, to over 2000 m in the 1-m precipitation per year zone of the eastern mountains (Chinn and Whitehouse, 1980).

Seasonal Snow - The Winter Snow Pack

In New Zealand, winter snow cover of the skifields and high peaks down to the bush-line normally commences accumulating in May, and lies through until October when it melts very rapidly. Of course the higher on a mountain, the earlier the first snowfall and the longer the snowpack lies, until at the regional snowline, the times of the last melt and first snowfall are effectively synchronous. The lowest elevation of the edge of the winter snow, the transient winter snow line, is naturally highly variable, and averages 1000 m in Southland and 1500 m on the North Island mountains. Typically snow in winter covers 35% of the South Island and less than 5% of the North Island (Figure 5.1).

The Formation of Snow

Snow forms in the atmosphere at altitudes of 2,000 to 12,000 m where ice and supercooled water coexist. Ice crystals commence growing, in a manner similar to rain droplets, until they are large enough to fall. The forming snow crystals are hexagonal in structure, but the traditional delicate stellar-shaped snow crystal is a rarity in maritime areas like New Zealand. Depending on the temperature and humidity at the time of formation, initial crystals take on many different forms like plates, rods, cups etc. As the crystals fall through humid air, they accrete ice rime which obscures the original crystal form. From the few observations of new snow forms in New Zealand (O'Loughlin 1969; Prowse 1981; Weir and Owens 1981), heavily rimed needles associated with humid conditions near the freezing level are the commonest form.

Growth of the Snow Pack

Maximum snowfall depths occur in the high precipitation zones along and west of the Main Divide. Nor'westerlies bring the largest masses of snow, but as the temperatures of this airflow are relatively high, nor'west snow tends to lie at higher elevations near the Divide and rarely reaches the ski-fields along the eastern margin of the foothills. It is the cold frontal southerlies that bring frequent but limited falls to lower elevations such as the eastern foothills. Our maritime climate means that air temperatures are rarely far from zero in the seasonal snow zone, and sporadic melt occurs throughout the winter.

The Wedge of Seasonal Snow

The frontal systems that bring the snowstorms have widely variable temperature structures, and freezing levels rise and fall over many hundreds of metres both within storms and between storms, so that the position of the snowline changes rapidly. Each successive storm deposits a snowpack that steadily increases in depth with height up the mountainside, giving a wedge shape to the profile of snowpack thickness (Figure 5.3). Snow that falls below the freezing level forms a wet snow zone (Figure 5.3a), while high on the mountains, where winter temperatures remain below zero, the snowpack is dry. Successive storm wedges accumulate to form a composite wedge-shaped snowpack (Figure 5.3b) whose shape changes throughout the year, and varies from winter to winter. In spring as freezing levels rise, the snowline retreats as the wedge thins, ultimately to disappear off the top of the mountain, or to survive to the next oncoming winter at the permanent snowline (Fitzharris, 1978).

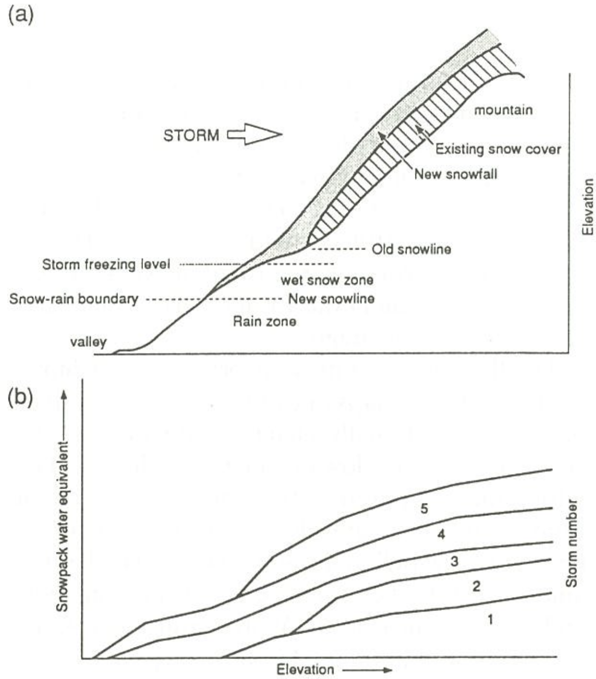

Figure 5.3 Development of the seasonal snow wedge.

Wind Redistribution

Snow is particularly susceptible to wind redistribution, and the snowpack wedge may be totally remoulded by wind. Snow is redeposited from windward slopes to accumulate in drifts in hollows and lee faces. Most redeposited snow accumulates just to the lee of ridge tops, while a small amount is blown to lower elevations where it melts. The highly variable pattern of snow redeposition is evident in the patterns of late-lying snow towards the end of spring. In a study of snow accumulation on the windswept mountains of central Otago, Harrison (1986) found that snow drifts formed a useful water resource, storing approximately 4 million m^3 in a 120 km^2 catchment, and contributing up to 9% of spring runoff. Across the ranges he found that slopes having south-east to west aspects recorded snowpacks with rainfall equivalents of less than 200 mm, while sheltered northeast slopes had accumulated over 450 mm.

Measurement and Estimation of Snow and Ice

Ski-fields lie in areas favourable to snow accumulation, and also provide a guaranteed winter access, so most snow studies have been made on ski-fields. Snow course studies have been made in the MacKenzie Basin at Ohau ski-field in the Barrier Range; Round Hill ski-field in the Two Thumb Range; in the Waimakariri basin at Alan's Basin and Camp Stream in the Craigieburn Range; and at Fraser basin in Otago.

Depth is the most obvious property of a snowpack, but it is perhaps one of the less useful values because it continually changes with time as the snow melts, packs down or metamorphoses. The information required is the equivalent depth of rainfall, or water equivalent. This value is not to be confused with liquid water content, which is the amount of water held, sponge-like, in melting snow and is rarely measured. Water equivalent is the amount of water obtained if the snowpack were to be melted. It is normally measured directly by weighing a core sample. Otherwise the water equivalent is simply the product of depth and density.

Snow density is the measurement used to convert snow depth to water equivalent. It is normally obtained by weighing a snow sample of known volume, and may be given in % or kg m^{-3}. For example, a snow density of 400 kg m^{-3} could also be expressed as 40%.

Manual Observations

Many different techniques have been used to measure snow and ice (Figure 5.4). The most straightforward is to measure snow depth at a point, but this is of little value without a density value. Precipitation gauges give the water equivalent of snow, provided that they are not capped or buried, and that the snow is melted before reading. The majority of snow measurements are made at a line of stakes on a snow course where the standard procedure is to make repeated measurements at almost the same locations. Usually the full depth of the snow pack is measured and sampled with a coring device, such as the Federal sampler (Figure 5.4d). This equipment is calibrated so that weighing the sample for a density measurement actually gives a direct reading of the water equivalent. Typical results of snow surveys are given in Figure 5.5 for four visits to the Round Hill course.

Glaciers present additional problems as there is no ground surface to use as a base datum. In addition the snowpack may be too deep or too hard to penetrate with the coring equipment. In these cases, the snowpack is measured in increments as it grows, and the snow surface is marked with a layer that can be found later in the core. Sawdust was used for a marker horizon in New Zealand glacier studies (Chinn, 1969).

Rates of snowmelt can be measured as the depletion of the snowpack water equivalent, or by the loss of snow depth measured against installed stakes. This latter method is used on glaciers, both in the snow of the accumulation area, and on the ice surface of the ablation area. Installing the stakes into glacier ice requires specialised drilling equipment. Bishop and Forsyth (1988) describe the use of a steam driven, hot point drill to place 10 m poles on the Dart glacier where ablation near the terminus averages 8 m yr^{-1}. Because high rates of ablation are characteristic of New Zealand glaciers, very long, flexible jointed ablation stakes must be used for all but short period observations. Ice densities of approximately 880 kg m^{-3} are often assumed in converting ice loss to water units.

Automatic Recording

More continuous information on snow deposition is obtained from recording precipitation gauges, though a turbulence damping device, such as an Alter shield (Figure 5.4e), is required to increase the accuracy of the catch. Snow depth variations may be obtained from an ultrasonic sensor (Figure 5.4f) while measurement of both accumulation and melt in water equivalent units can be made with a snow pillow (Figure 5.4g) in which the pressure of the snow pack on a reservoir bladder of non-freezing liquid is recorded.

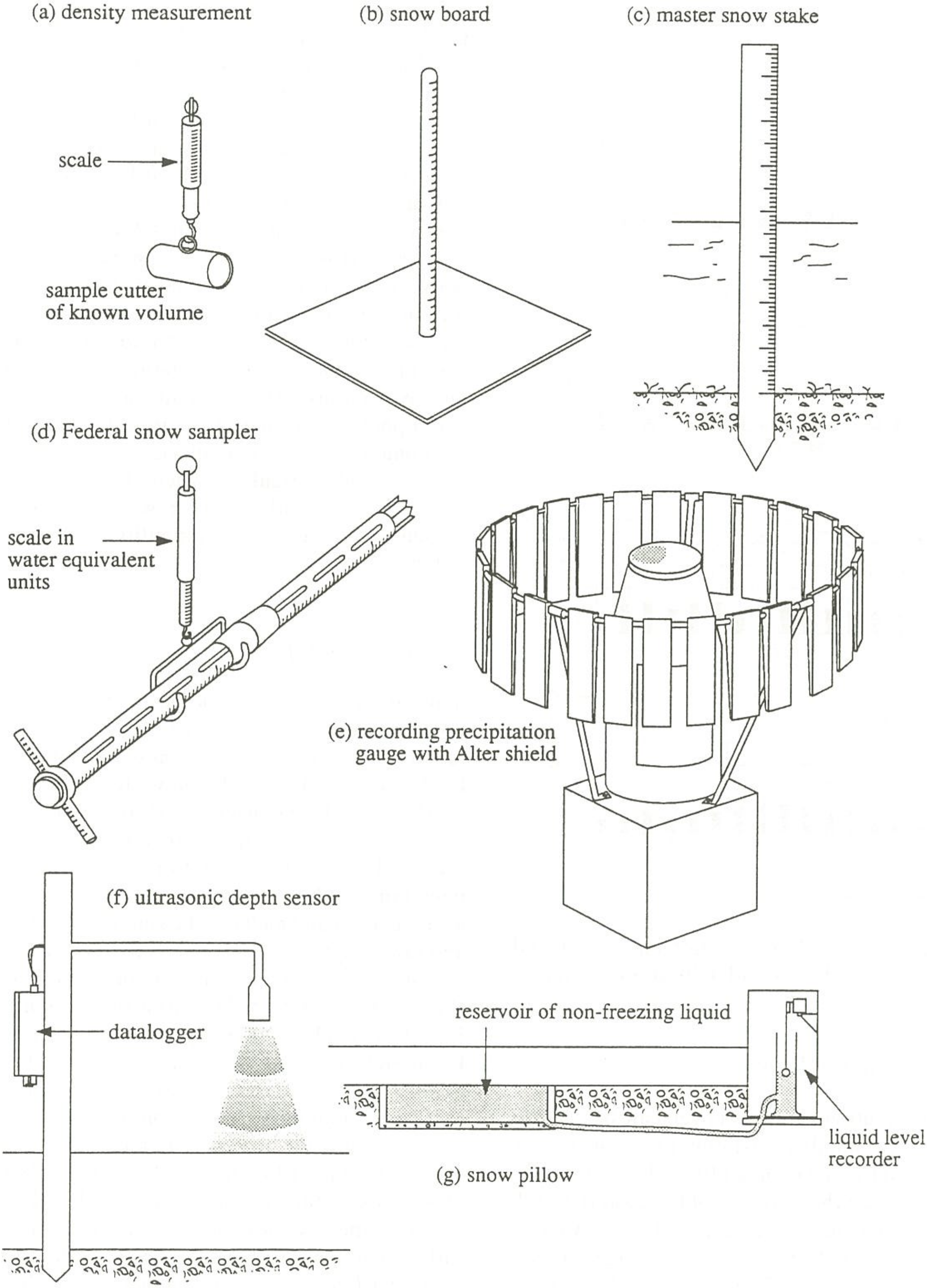

Figure 5.4 Instruments for measuring snow accumulation and melt.

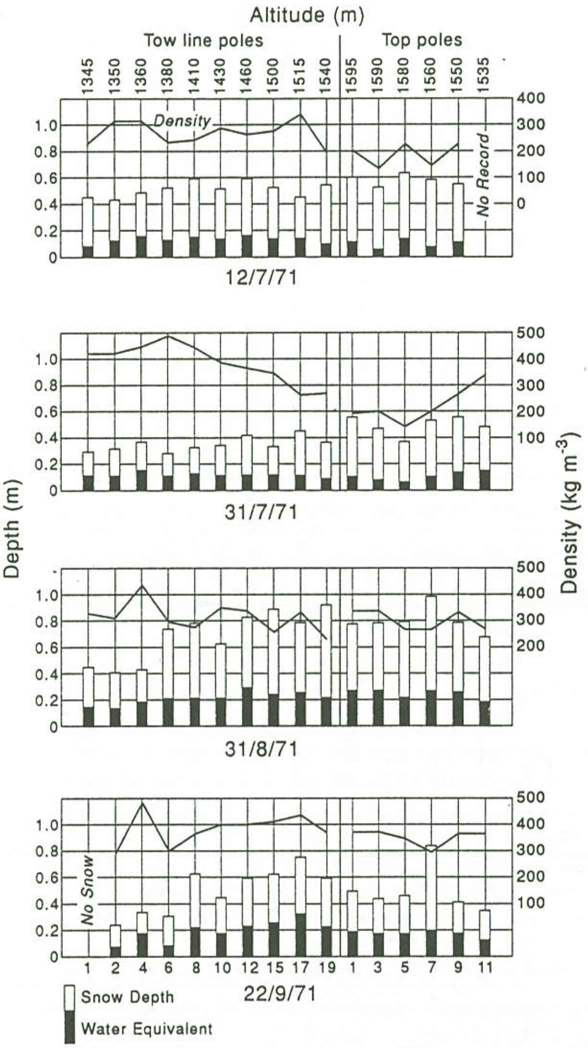

Figure 5.5 Snow depths, water equivalent and densities from the Round Hill snow course in 1971.

Remote Sensing of Snow

Areal distribution of snow cover can be plotted from a large number of point observations, as Chinn (1981) and Hughes (1974) have done for storms affecting the lowlands of the eastern South Island. More commonly, the extent of snow cover is assessed remotely from satellites using radiation sensors. For example, Thomas et al. (1974) used Landsat imagery for snow investigations on the St Arnaud Range. Although these images have good resolution, they are of limited use for hydrological applications because they do not give snow depths, reading intervals are limited to 18-day orbit intervals, and cloud cover frequently obscures the scene. Weather satellites have much lower resolution, but images are available at least daily. The potential for their use has been examined by Hickman (1972) and Fitzharris and McGann (1989), who show that there was reasonable agreement between transient snow line estimates from satellite imagery and from weather stations. Aerial photographs can be used to identify glacier snowlines at the end of the ablation season (Chinn and Whitehouse 1980). Landsat imagery and aerial photography have been useful for inventory mapping of New Zealand glaciers, although there are frequent difficulties when the glaciers are obscured by cloud cover or winter snow, and resolution is insufficient to identify small glaciers (Chinn 1989a).

Calculated Snowpacks

Snow accumulation or melt may be fairly successfully estimated from standard meteorological observations for areas with no snow measurements. Fitzharris (1987) created a snow-storage index (S) for the South Island using weekly departures from the average of precipitation and temperature recorded at 21 climate stations adjacent to the mountains. Alternatively, the processes of snow accumulation and melt may be simulated as Moore and Owens (1984a) did for the Alan's Basin snow course at 1750 m elevation in the Craigieburn Range. Precipitation data from the Craigieburn Forest site (914 m) and temperature data from Ski Basin site (1550 m) were used. The model required optimization of parameters for the rain/snow boundary, a precipitation correction factor to allow for increases with elevation, and a degree-day factor for snow melt. More elaborate definitions of the rain/snow threshold and more detailed specification of temperature variation with height are possible to improve such models. Barringer (1989) simulated snowline elevation on the northfacing slope of the Remarkables Range

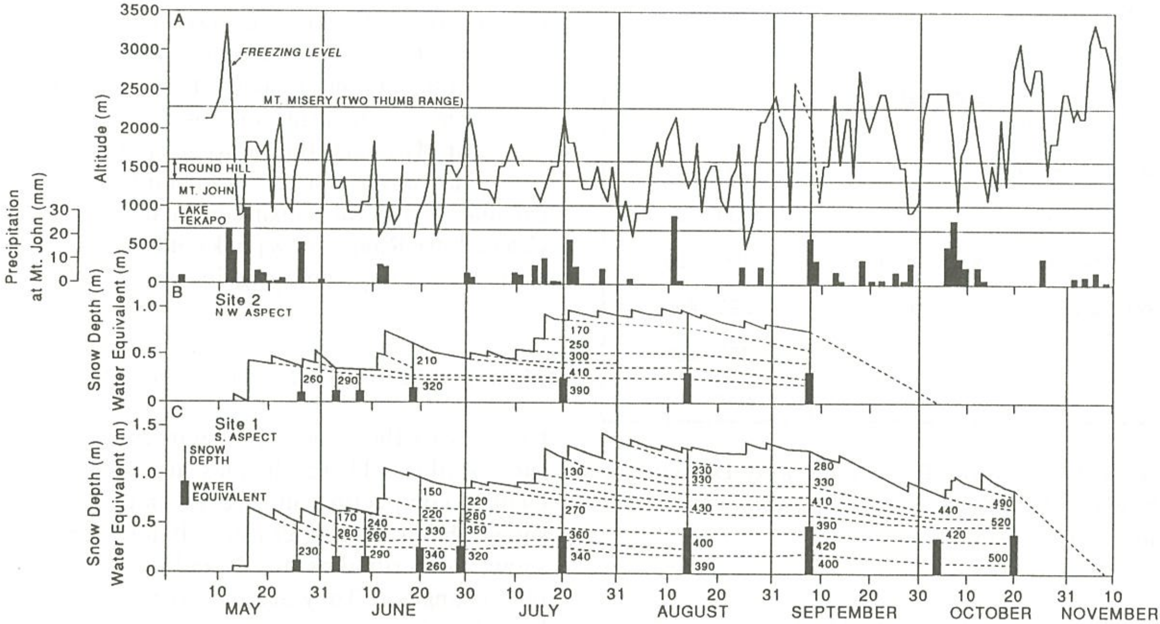

Figure 5.6 Detailed growth and decline of the 1972 snowpack at Round Hill snow course. Densities given in kg m^{-3}.

using data from the meteorological station at Queenstown airport and automatic recording stations at 950 m, 1295 m and 1615 m elevations.

Structure of the Snowpack

Snow Depth

The dates when the snowpack begins to accumulate, reaches a peak depth, and finally disappears depend on altitude. The higher the elevation, the earlier the first snow of the season, the thicker the maximum pack and the longer the pack lies. Snow depths at high altitudes do not reach their maximum until well after the "ski season" at medium altitudes has finished. The evolution of a snowpack is available from detailed measurements made in two pit sites dug for each survey at Round Hill (Figure 5.6). Each snowfall is related to storm precipitation (measured a few tens of kilometres away at Mt. John) and to the freezing level. Towards October, when the freezing level rises, few additions were made to the snowpack, as precipitation fell mostly as rain. Each snowfall makes a step increment to the snowpack depth, which decreases exponentially with compaction and metamorphism.

Snow Density

New snowpack densities increase as temperature of the falling snow increases, so that snowfalls at higher, colder elevations tend to be lighter. However wind-packing and rimed snow ensure that most of our high altitude snow also forms high density packs (Prowse and Owens 1984). The common belief that snow is only 10% water is a myth in New Zealand - the average snowpack density is normally close to 300 kg m^{-3} (Chinn 1981). Almost invariably our snow is "wet" and forms packs of around 150 to 250 kg m^{-3} early in winter which steadily increases, with thickness and metamorphism, to 350 to 450 kg m^{-3} by the end of winter. In packs over 1 to 2 m in depth, densities

Type	Density (kg m-3)
Wild snow	10 - 30
New snow, immediately after falling in calm	50 - 70
Damp new snow	100 - 200
Settled snow	200 - 300
Wind packed snow - soft slab	100 - 290
Wind packed snow - hard slab	290 - 450
Firn	400 - 800
Glacier ice	917

Table 5.1 Snow-Ice Types and Density of Snow and Ice. Source: after Seligman (1936) and Paterson (1969).

of 500 kg m^{-3} may be reached by late spring. Typical densities are given in Table 5.1 and in vertical profiles measured throughout the winter in Figure 5.6, together with water equivalent depths.

The densities of snowpacks on glaciers show a greater range of values because of the greater pack thickness, and the densities continue to increase downward through the snow that has survived from previous winters (known as firn). Near-surface densities commence at around 300 kg m^{-3}, and increase to 500 kg m^{-3} at the base of the recent winter's pack. Thereafter the density profile increases uniformly through three or four years of firn to the 880 kg m^{-3} density of glacier ice (less than the 917 kg m^{-3} of pure ice because of some included air).

Temperature Regime

The snowpack can be at any temperature up to 0°C, and it is normally well below this temperature when forming in early winter. Although the vertical distribution of temperature within the snowpack and its changes as winter proceeds can be quite complex, the general trend is for a progressive warming until the snowpack becomes isothermal at 0°C. The snow is then described as ripe (ripe for melt) and contains liquid water. In our maritime climate, the lower elevation snowpacks may become isothermal at any time.

The cold content, or heat deficit, of a snow pack below zero is a measure of how much energy will be required to warm the snowpack until it becomes ripe. Our snow is not very cold and Prowse (1981) calculated that the maximum heat deficit of the Craigieburn Range snow pack could easily be overcome by one day's energy gain, even in winter.

Metamorphism - the Evolution of the Snowpack

Changes to the snowpack begin the instant the snow settles. These changes are by vapour exchange, compaction and freeze and thaw in ripe snow. Even at temperatures below zero, when snow is described as dry, it undergoes metamorphic changes. They are of two main forms and have been studied extensively in avalanche research.

Equilibrium growth (formerly called equi-temperature) metamorphism occurs where temperatures throughout the snowpack are nearly uniform. The mechanism is by 'vapour exchange' from sharp corners and small crystals to produce rounded snow grains which are well bonded together. The resulting snowpack is very stable. This type of metamorphism is common in maritime areas with deep, relatively warm snowpacks. Consequently it is a dominant process in New Zealand snow.

Kinetic growth (formerly temperature-gradient) metamorphism should be of particular interest to skiers because it is this process, hidden beneath the surface, that surreptitiously manufactures avalanche conditions. This type of metamorphism is driven by strong vertical temperature differences within the snowpack. Vapour transfer initially forms faceted crystals structures, similar to frost which eventually grow into delicate hollow cup-shaped crystals called depth hoar. Left alone a weak depth hoar layer may avalanche itself, but if the slopes are loaded with a heavy new snowfall, the chances of an avalanche are greatly enhanced. Although depth hoar is characteristic of cold inland locations with shallow snow packs, it is surprisingly common in New Zealand, despite our maritime climate. It is observed in colder winters in the eastern ranges where snow packs are shallower and cold clear conditions may persist for long periods (McNulty and Fitzharris 1980, Prowse and Owens 1984, Weir and Owens 1981).

When liquid water is present in snow, metamorphism occurs very rapidly and produces clusters of grains which eventually lose their identity to form large loose snow grains known to skiers as spring corn. Re-freezing of layers following melting may form very strong and thick ice crusts.

Processes of Snow and Ice Melt

Melting of snow and ice is a complex process which depends on the sources of energy supplied, all of which vary with the weather and the season. The most important sources of melt energy (Q_M) are radiation (Q^*), sensible heat (Q_H) and latent heat (Q_E) (Figure 5.7). Radiation energy is gained as solar (short wave) and long wave radiation from surrounding ground and sky and is lost by short wave reflection and upward emission of long wave radiation to space. Bright new snow may have a reflectivity, or albedo, as high as 95% so that only

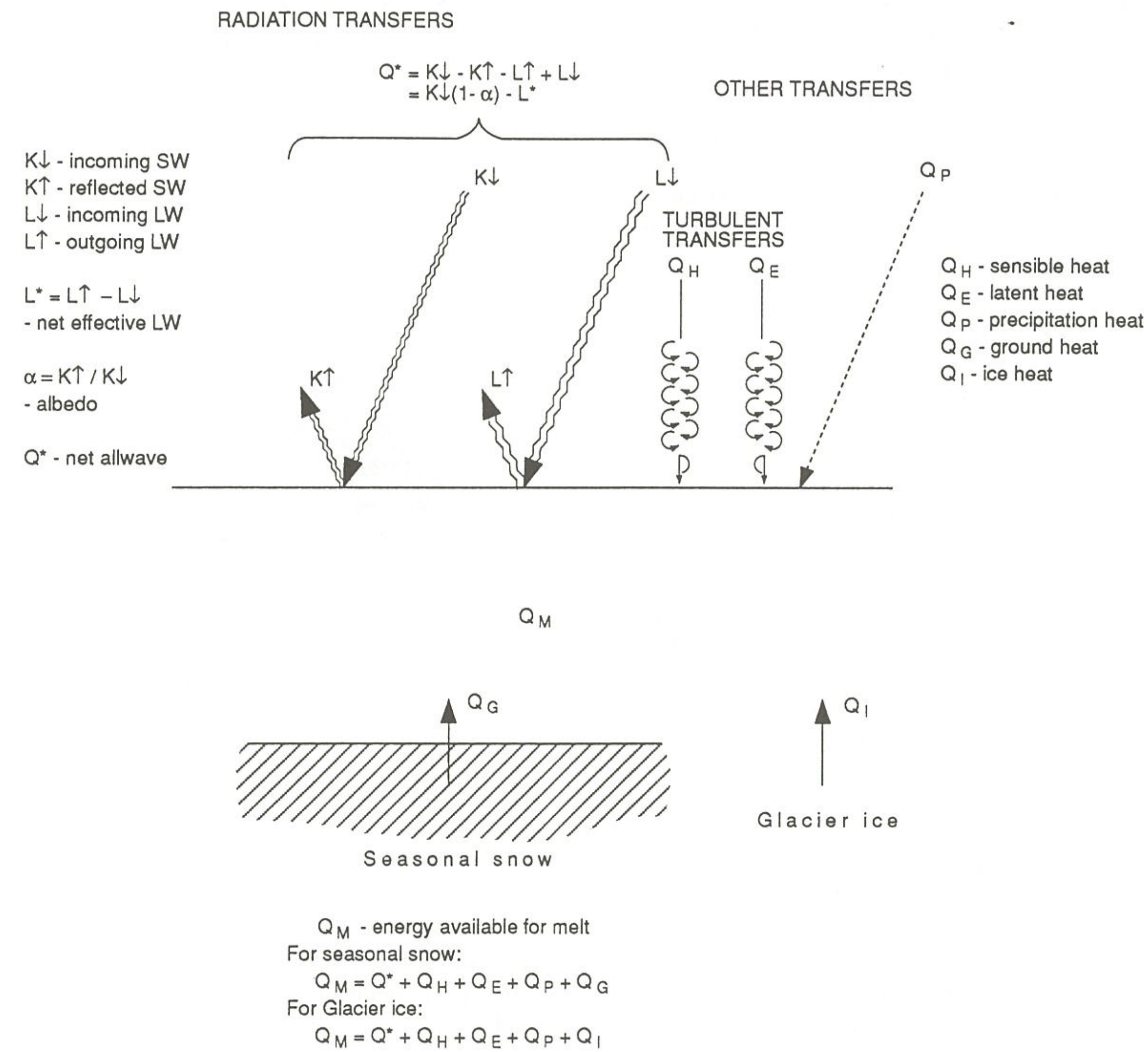

Figure 5.7 Sources of energy for snow and ice melt.

5% of the solar energy is absorbed, but for melting snow and ice surfaces, albedoes are usually closer to 50%.

Sensible and latent heat energy amounts depend respectively on the temperature and humidity of the air overlying the snow or ice surface, as well as the wind speed which enhances turbulent motion of the air. In the case of latent heat, energy is gained as humid air condenses on the snow surface.

The widespread belief that the energy from precipitation (Q_P) causes rapid melt, even with small amounts of warm rain, is not strictly true. It probably arises from the observation that snow depth decreases rapidly during rain. Often, however, this results from gains in density or is associated with large turbulent energy gains as temperature and humidity are characteristically high in windy nor'west rainstorms in the mountains. Energy transfers by conduction from the ground (Q_G) or deep glacier ice (Q_I) are generally small, and are often neglected.

Relative Amounts of Energy

Studies made around the world show that radiation is usually the main cause of seasonal snow melt, especially in forested locations and at high elevations. Under New Zealand's maritime climate, however, turbulent transfers are more important; that is, the heat is mainly supplied by warm, humid winds. Energy from rain (Q_P) also makes measurable contributions. Studies undertaken during rapid melt in central Otago, the Craigieburns and at Arthur's Pass show that the turbulent transfers and rain contribute 70-80% of the total melt energy (Fitzharris et al. 1980, Prowse and Owens 1982, and Moore and Owens 1984b) during flood events.

Major heat balance studies have been made on two New Zealand glaciers, the Ivory and the Franz Josef. The Ivory Glacier, in the headwaters of the Waitaha River in Westland, was studied in detail as an International Hydrological Decade Basin. Heat balance studies were made at an elevation of 1450 m during two summer ablation seasons (Hay and Fitzharris 1988). The average results of the intensive measurements (Figure 5.8a) show that energy gains were dominated by radiation. The lower trunk of the Franz Josef Glacier in South Westland has been the site for a series of short-period energy balance studies which cover most seasons (Marcus et al. 1985, Ishikawa et al 1992). Average energy gains for the comparatively low

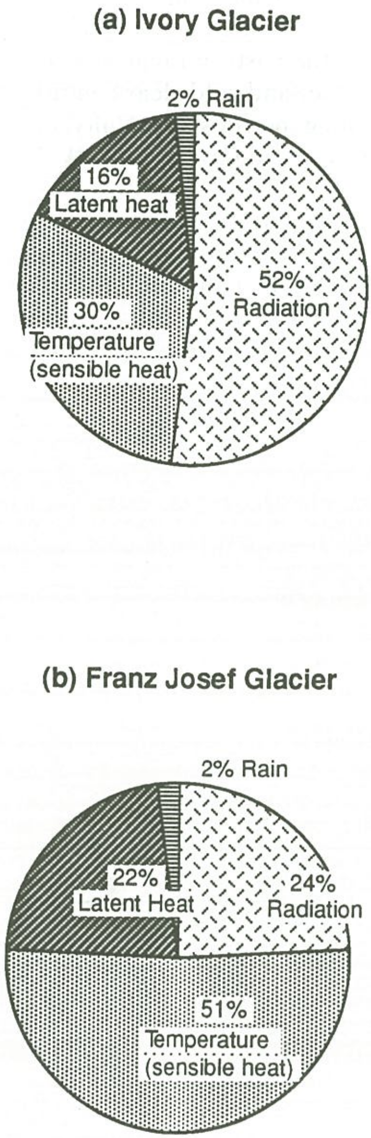

Figure 5.8 Relative sources of energy for ablation losses on the Ivory and Franz Josef Glaciers.

500 m elevation site are shown in Figure 5.8b. Here the warmer temperatures of the lower altitude dominate so that sensible heat energy is most important, and despite rainfalls comparable with any in the world, precipitation accounts for a meagre 2% of the melt.

Meltwater Runoff

The runoff contributed by snowmelt is generally around 12% of the annual discharge of New Zealand's larger alpine rivers. In some smaller catchments, seasonal snow melt makes greater contributions. For example, in the Fraser catchment of Central Otago where runoff is used for local hydroelectricity generation, irrigation of horticultural land, and for frost fighting, snow storage amounts to 33% of the annual river flow (Fitzharris and Grimmond 1982).

New Zealand's main hydro-electric catchments lie to the east of the Southern Alps where spill-over precipitation from norwest storms (Salinger 1979, 1980) provides the bulk of the runoff, with negligible additions from occasional southerly and easterly storms. A total area of 26,000 km^2 of South Island mountain catchments supply inflow to hydroelectric storage lakes, and within these catchments are large areas of permanent snow and ice. Seasonal snow storage in these catchments is equivalent to 330 mm water equivalent over the total area, or 15% of the annual runoff (Table 5.2). This storage represents water that is guaranteed to appear as river flow over the period October to January. It is almost the same size as the 17% of annual runoff which is harnessed as controlled lake storage. Both of these storages have a very high monetary value as potential hydroelectric output.

Variability of Snow and Ice Storage

Annual differences in Snowpack Size

The size of seasonal snow storage can vary markedly from winter to winter. Unfortunately, there are only two snow courses in New Zealand with sufficiently long records to estimate year-to-year

Parameter	Water units (mm)	Energy units (GWh)
Precipitation	2 820	
Evaporation	550	
Runoff	2 270	15 550
Seasonal snow storage	330	2 260
Controlled lake storage	394	2 700

Table 5.2 Estimated long-term water balance for the combined South Island hydro-electric catchments. Source: modified from Fitzharris (1987)

variability in seasonal snow storage. At Alan's Basin (elevation 1750 m) in the Craigieburn Range, peak snow accumulation varied between 230 mm and 1030 mm over a 12-year period, with a standard deviation of 240 mm and coefficient of variation of 44% (Morris and O'Loughlin 1965). A 19-year record of snow pack variations from Round Hill in the Two Thumb Range showed a coefficient of variation of 52% (Chinn 1981). In the Fraser catchment (120 km^2), Harrison (1986) reports the maximum volume of water stored as seasonal snow was 23.3 million m^3, but over a period of 17 years some winters recorded only about half of this. Certainly the seasonal snow line varies greatly both within a winter and from year to year, although at least on the Remarkables near Queenstown there appears to have been little trend in the mean value over the last five decades (Barringer 1986, 1989). Variations over eight winters in the snow storage index of Fitzharris (1987) are shown in Figure 5.9. The index appears to be sensitive, and suggests snow storage was substantially greater in the winters 1980, 1981, 1983, and 1986, but smaller in 1984 and 1985. The winters 1979 and 1982 were about average.

Snow for Hydroelectric Generation

When variations in winter freezing levels and precipitation are analysed for a 25-year period, it

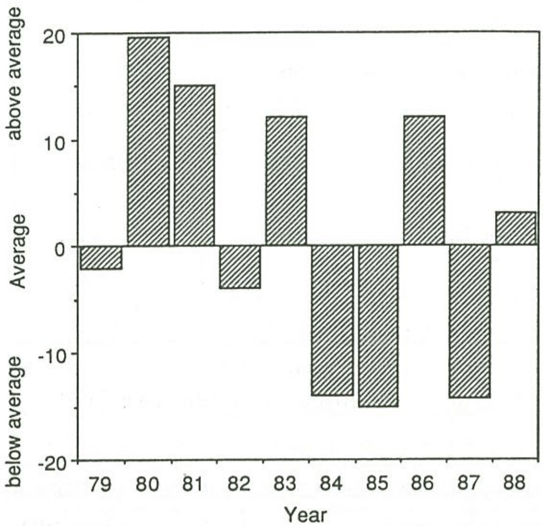

Figure 5.9 Behaviour of snow storage index (shown on vertical axis) over eight winters. The horizontal line at S = 0 indicates average conditions.

appears that seasonal snow storage in an averaged South Island hydroelectric catchment can vary between 120-1160 mm, or 440-4370 GWh. This has substantial economic implications as it amounts to a range of about $15-$160 million (in 1992 terms). Spring runoff over the period 1931-84 has varied by ±500 mm about a mean of 800 mm. Fluctuations in snow storage have accounted for about 40% of this variation, while fluctuations in spring precipitation have accounted for 60% (Fitzharris 1987). All these results confirm that the amount of water stored as snow can be markedly different from one winter to the next, and as a consequence spring snow melt runoff will never be consistent.

Influence of Snow and Glaciers on Flow of Alpine Rivers

Seasonal patterns of runoff show increasing variation in river flow with increasing proportion of the catchment area under permanent and seasonal snow. Contrasting types of distribution are illustrated in Figure 5.10. For the glaciated Hooker catchment, maximum flows occur in January and this month alone accounts for almost 20% of the annual total. There follows a steady decrease in runoff to the winter months, when most precipitation falls as snow and there is little melt. Thus July delivers only 2% of the annual flow. River flows rise again with spring melt, closely following the insolation cycle and availability of energy for melt.

By contrast, the Manuherikia catchment in Central Otago has no permanent ice, but is largely covered by seasonal snow in winter. Runoff peaks in October as temperatures rise and the bulk of the seasonal snow is lost. After most of the snow has melted, the flow diminishes rapidly to a minimum in February, and is only sustained by rare rainfalls, melt from remnant snow patches and base flow from upland bogs and groundwater. Catchments with no snow or ice, such as the Motu, tend to

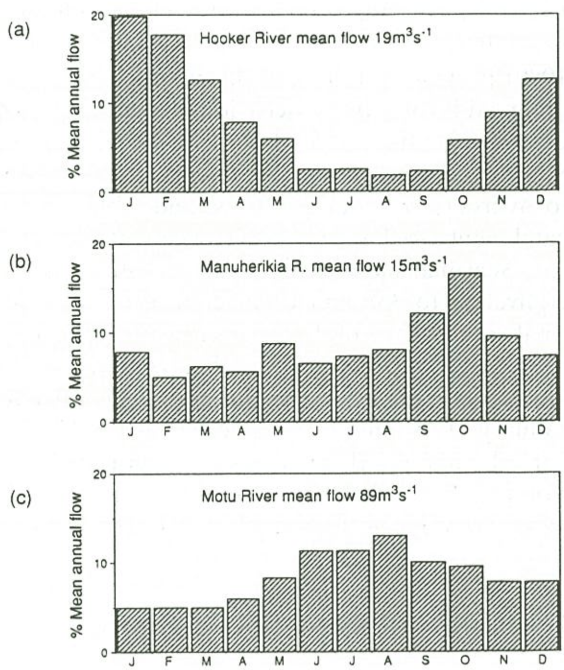

Figure 5.10 Mean monthly runoff for contrasting catchments showing the role of (a) glaciers, (b) seasonal snow and (c) no ice or snow.

produce a more even flow distribution throughout the year, with any variation produced by seasonal changes in evaporation and rainfall.

The Fraser River is a high catchment with pronounced winter snow accumulation. Diurnal variations in river flow from daily melt are clearly defined as spikes superimposed on an upward bulge in spring base flow (Figure 5.11). The size of the spike for any given day depends on the synoptic weather pattern and the energy available for snow melt. On anticyclonic days there is much solar radiation for melt and the spikes are pronounced. They grow even larger with the advent of warm nor'westers ahead of a cold front, when sensible heat flux is high, and all but disappear during cloudy, cold southerlies that follow a frontal passage.

Since the end of last century our receding glaciers have been releasing water from long-term storage, in excess of the yearly input from snowfall. Retreat is continuing, with intermittent small re-advances, analogous to waves on the beach when the tide is going out. This water makes a significant contribution to the flow of the Waitaki River. If the estimated average net mass loss for the Tasman Glacier over this century of about 1 m water (Goldthwait and McKellar, 1962; Skinner, 1964) is assumed to apply to all glaciers in the Lake Pukaki catchment, then the mean annual contribution of water from long-term ice storage is about 190 million m^3. If this melt water is released from January to March, then the contribution is equivalent to a mean flow of 24 m^3s-1, and represents about 10% of late summer inflow to the lake (Anderton 1973). The contribution is greater in warmer summers, and helps offset low flows from non-glaciated hydro catchments in dry seasons.

The only other analysis of the contribution of glacier melt to river flow comes from the Ivory Glacier. Here snow forms about 25% of the annual precipitation and its melt contributes 21% of the runoff. About 9% comes from melt of perennial snow and ice, although the relative proportions vary markedly from year to year (Anderton and Chinn 1978).

Snow and Ice Floods

Under some conditions, rapid snow and ice melt may be an important contributor to floods. The role of snow melt during a major flood in the Clutha River has been assessed for two sub-catchments by Fitzharris et al. (1980). In the Fraser, where 80% of the catchment was above snow line, melt provided 40% of flood runoff. Melt rates of nearly 3 mm hr^{-1} preceded rainfall. These saturated and primed the catchment, so that very quick rises in streamflow occurred when 100 mm of rain fell. In the Pomahaka, only 10% was above

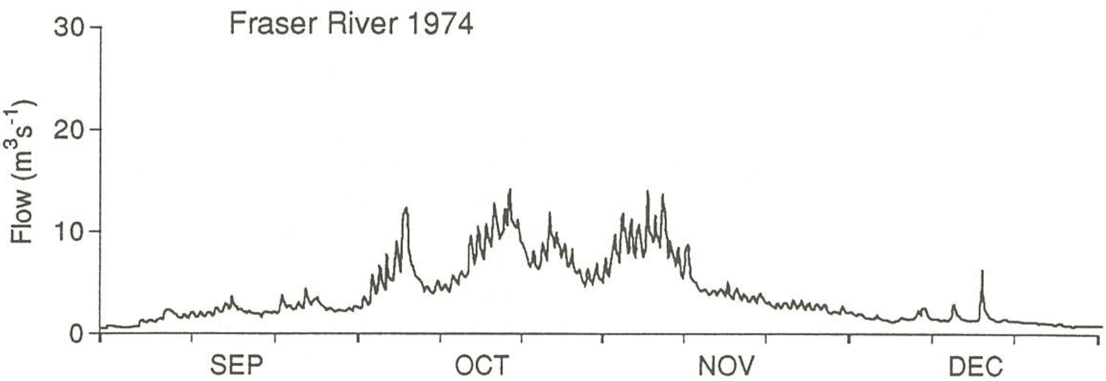

Figure 5.11 Typical pattern of daily spring flow for the Fraser River, Central Otago.

snow line, so snowmelt provided only 10 mm of the estimated 66 mm storm flow. Thus the relative contribution of snow melt to floods depends upon snow line elevation and snow covered area, as determined by the catchment hypsometric curve.

Accelerated melting of transient snow at lower than normal snow elevations may contribute to small floods, because when the snow line is low a large proportion of the catchment contributes melt. For instance, Moore and Prowse (1988) described a storm which caused road washouts in the Craigieburn area in July 1983. Precipitation of 123 mm in one day was augmented by over 25% by melting of a shallow snow pack at relatively low elevations. More rarely, floods can arise from snowmelt alone. For example, the largest daily flow of Camp Stream (Craigieburn Range) for the whole of 1982 occurred on November 7th from strong melt associated with dry weather and foehn conditions (Moore and Prowse 1988).

Rapid ice melt on glaciers seldom leads to flooding, because only relatively small areas of ice extend to low elevations. However, sudden releases of water bodies dammed alongside, within, or on glaciers, called "jokulhlaups", periodically occur. For example, they have been reported from the Franz Josef Glacier in 1965, 1981 and 1990 (Brazier et al. 1992). The most devastating event of this kind, however, led to the Tangiwai disaster of 1953. Sudden drainage of glacier-dammed Crater Lake on Mt Ruapehu initiated a lahar, which demolished a bridge on the main trunk railway, and over 150 people perished. Similar rapid drainage of the lake occurred about 1860, 1895 and 1925 (O'Shea 1954).

Effect of Past and Future Changes of Climate

Glacial Retreat

Our glaciers have been dwindling in volume since their last maximum extents about 1860 to 1890 when they were first sighted by European explorers. New Zealand has been warming since about 1900 (Salinger 1976, Royal Society 1989), and a significant positive correlation has been found between the retreat of the Dart and Stocking Glaciers and temperature (Salinger et al 1983). The Dart glacier has shrunk and retreated over 5 km since 1850 (Bishop 1977, Bishop and Forsyth 1988).

On the western side of the Main Divide, precipitation changes appear to affect glaciers more than temperature changes (Suggate 1950). The detailed record of variations in the terminus of Franz Josef Glacier are significantly correlated with precipitation measured at the nearby township and at Hokitika, with a lag time of 5 years (Hessell 1983). During the mid 1980s all glaciers had positive balances, apparently due to heavy precipitation produced by a series of El Niño events which often bring sustained westerly airflows over the Southern Alps. This pulse of increased snow has passed through those glaciers with a high activity to produce an advance of the glacier front. In the case of the Franz Josef, the terminus moved forward 600 m.

Our glaciers are sensitive to the frequency of synoptic weather types in the South West Pacific region, especially the relative strengths of the westerlies and of blocking anticyclones (Hay and Fitzharris 1988). Thus one possible reason for the general retreat of Southern Alps' glaciers is that there has been an increase in the frequency of synoptic weather types that favour high ablation rates.

Although the relationship between glaciers and climate is complex, they are one of our better indicators of climate change. If the predicted 1.5 $^{\circ}$C to 4.5 $^{\circ}$C of global warming were to occur from a doubling of greenhouse gas concentrations, most small New Zealand glaciers would not survive, despite possible increases in precipitation. The higher temperatures would lift the permanent snowline above more than one third of our 3000 glaciers, and they would disappear within a few decades (Chinn 1989b). Large glaciers, such as the Tasman, respond very slowly to climate change and are at present out of equilibrium with the present climate. Because of their great ice mass, they will continue to exist for several centuries, even with global warming. As ablation quickens, their contribution to river flows will accelerate in importance.

Our very large glaciers might survive, but what of seasonal snow? Assuming increases in temperature of 3°C and precipitation of 15%, Fitzharris (1989) suggests a rise in snow line of 300-400 m, a decrease in snow accumulation below 2300 m and a reduction in winter snow-covered area for South Island hydro catchments from 45% to 28%. These changes will markedly alter the flow regime of many South Island rivers. Modelling studies suggest an increase in inflow to hydro storage lakes of 40% in winter, but a decrease of 13% in summer. Annual runoff would increase by 14%. All these changes favour increased hydro generation and reduce the demand for water storage. Such forecasts are preliminary, however, because current year-to-year variability in snow storage is larger than the predicted changes in the mean, and because of the uncertainties surrounding any regional estimate of future climate.

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6
Analysis of Floods and Low Flows

C P Pearson

Introduction

At the end of January in 1984, much of Invercargill and the towns of Otautau and Tuatapere lay under water, in the worst flooding in New Zealand's recent history. That flood was just one of many which, in the decade 1976-85, claimed seventeen lives and cost over $600 million (indexed to 1985 values). In 1988, a single storm, Cyclone Bola, alone caused an estimated $90 million of damage, particularly in the East Cape region. At the opposite extreme, droughts can be equally expensive; the Canterbury drought of 1987-8 cost the community an estimated $360 million.

Economic losses caused by flooded homes and businesses, lost agricultural production, damage to flooded pastures and farm buildings, disrupted transport, and so on are not the only effects of floods and droughts. The costs of environmental damage and social disruption are incalculable, such as those caused by soil erosion during heavy rain, the loss of rural confidence and consequent social changes which result from a major flood or drought, or restrictions on water use or power generation during droughts.

The devastating effects of floods have led to a variety of measures being taken to safeguard lives and property. These include measures such as construction of stop banks, floodways, and detention dams, waterproofing or elevation of buildings (termed structural measures) and preparatory and regulatory measures such as flood plain zoning and building restrictions, flood warning systems, civil defence measures, insurance, and so forth. The structural measures have tended to predominate, perhaps reflecting the long history of engineering approaches to river control in New Zealand, but the need to place more emphasis on non-structural measures has in recent years been strongly advocated (Ericksen, 1986).

Measures to anticipate and mitigate the effects of drought and low river flows have received much less attention, as the disarray in response to declining hydroelectricity generating capacity during the winter of 1992 graphically demonstrated. Nevertheless, as New Zealanders become increasingly aware that water resources are abundant in only some areas - which are mostly remote from the centres of greatest demand - a rapidly growing emphasis on the analysis of droughts and low flows can be expected.

New Zealand's weather is highly variable, and so too is the occurrence of floods and droughts. During the 1980's, there were serious floods in many parts of the country, including Invercargill, Greymouth, Taranaki, and Gisborne. Potentially serious floods were averted in other places by the stopbanks which have been constructed along thousands of kilometres of river. Drought affected eastern parts of both islands during 1987-88, but during the winter of 1992 affected virtually the entire country, with particularly serious consequences for the electricity generation industry.

This chapter considers floods, and droughts as they are experienced as low river flows. Droughts may be defined as periods of water deficit in one or more components of the hydrological cycle, but are ultimately caused by sustained periods of low or zero rainfall.

Floods and Droughts As Fluctuations In Streamflow

Variations in weather patterns are reflected in fluctuating river flows. The peaks in a river's flow, such as those shown on the hydrograph for the Hutt River in Wellington (Figure 6.1), are responses to rainfall and snowmelt. Flows rise sharply in response to rainfall, and decrease more slowly as the soil and rocks in the catchment gradually release the water that was stored during the storm. The flows of many rivers in New Zealand are modified by large lakes, which tend to reduce the size of floods and maintain flows during rainless periods.

The magnitude of a flood depends principally on the amount and duration of rainfall, and on the catchment's state of wetness beforehand. The extent to which flows decline depends on the period of time to the next rainfall, and on the amount of water stored in the catchment in the form of groundwater, soil moisture, snow, and in lake storage.

The principal questions asked about floods and low flows are "How severe, frequent, and prolonged are they?", and "When will the next one happen?" The first question is asked when the engineer or water resource manager must design a structure like a stopbank or irrigation scheme, or prepare a water management plan. The second is asked when preparations must be made to mitigate the immediate effects of a flood or drought, for example by evacuating stock from flood plains or scheduling thermal electricity generation to conserve hydroelectricity generating capacity.

The variability of river flows can be illustrated by a flow-duration curve (Figure 6.2). A flow-duration curve shows the percentage of time that flow equals or exceeds a particular amount, and hence provides a picture of the proportion of the time that a particular river experiences high, medium, or low flows. The slope of a flow-duration curve relates directly to the variability of a stream's flow . Different rivers can be compared

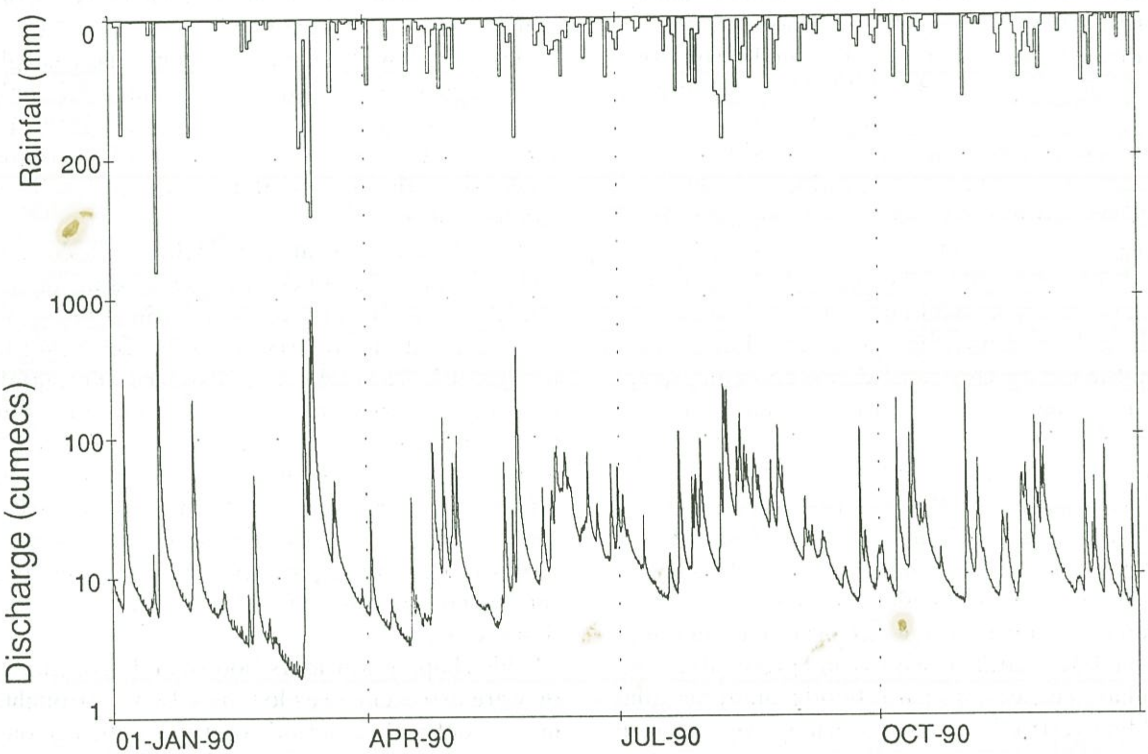
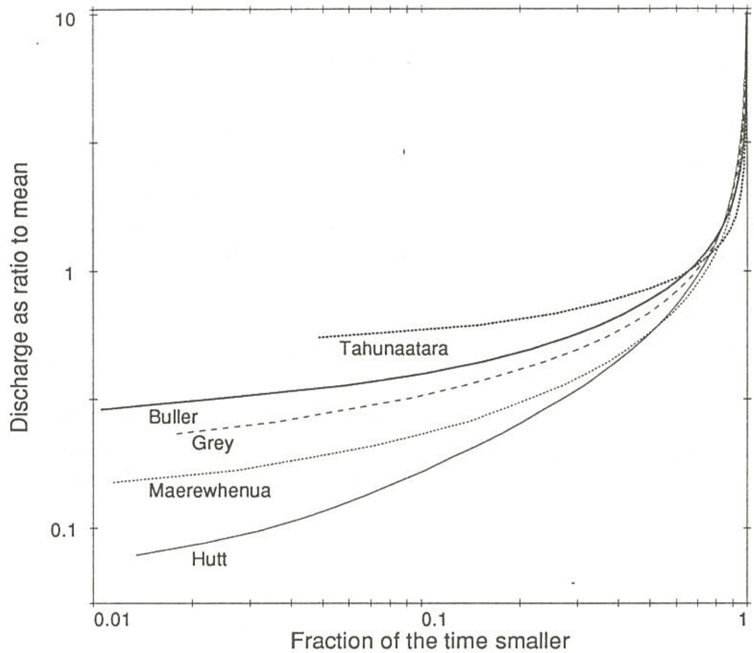


Figure 6.1 1990 hydrograph for the Hutt River at Birchville, Wellington (site number 29818, drainage area 427 km^2, Walter 1990). Daily rainfall totals from Phillips pluviograph (located inside Hutt Catchment) are shown for the same period.



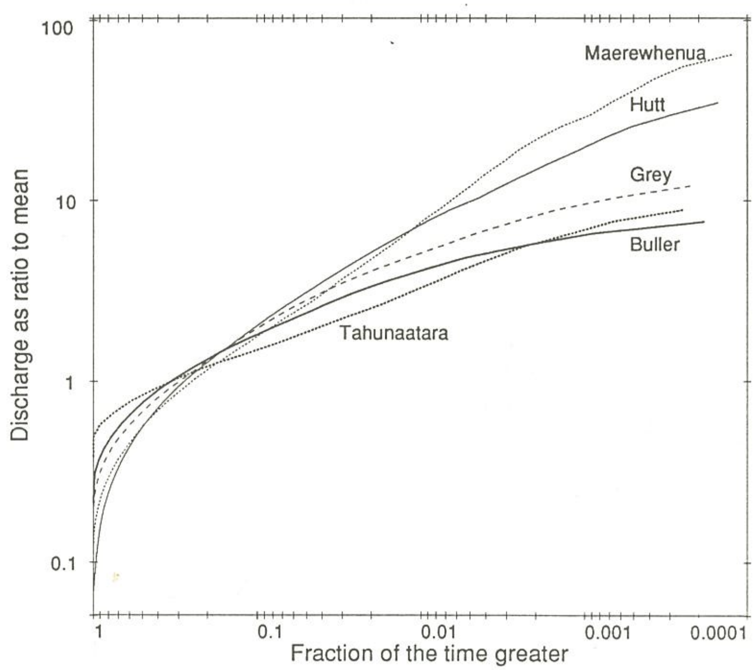


Figure 6.2 Comparison of flow-duration curves for five New Zealand rivers (site numbers in brackets) over the 20-year period 1972-91: Grey (91401, West Coast), Buller (93216, Lake Rotoiti Outlet, West Coast), Hutt (29818, Wellington), Tahunaatara (1043428, central North Island) and Maerewhenua (71106, South Canterbury).

by plotting their range of discharges as a ratio of their average discharge. For example, the Buller, Grey, Hutt, Maerewhenua and Tahunaatara rivers are compared in Figure 6.2. The Tahunaatara, with its porous basin, the Buller, in a high rainfall area with a lake outlet, and the Grey, on the wet West Coast have lower variability, while the Hutt with an impermeable basin and the Maerewhenua, in dry North Otago, are much more variable. The upper plot shows Tahunaatara and Buller low flows are well sustained whereas the Hutt River is faster draining. The lower plot shows Maerewhenua floods can occasionally be very large whereas Grey, Buller and Tahunaatara floods are usually less than ten times mean flow.

Floods

A flood is a discharge in a river or stream which exceeds the capacity of the channel and inundates neighbouring areas of normally dry land. Floods are commonly caused by heavy rainfall, but may also result from rapid snowmelt during warm weather (especially with warm rainfall), or from more unusual occurrences such as the collapse of a dam.

Flood Frequency Analysis

Floods may be described by reference to their peak discharge or water level, the time to peak, and the volume of storm runoff (Figure 6.3). A commonly

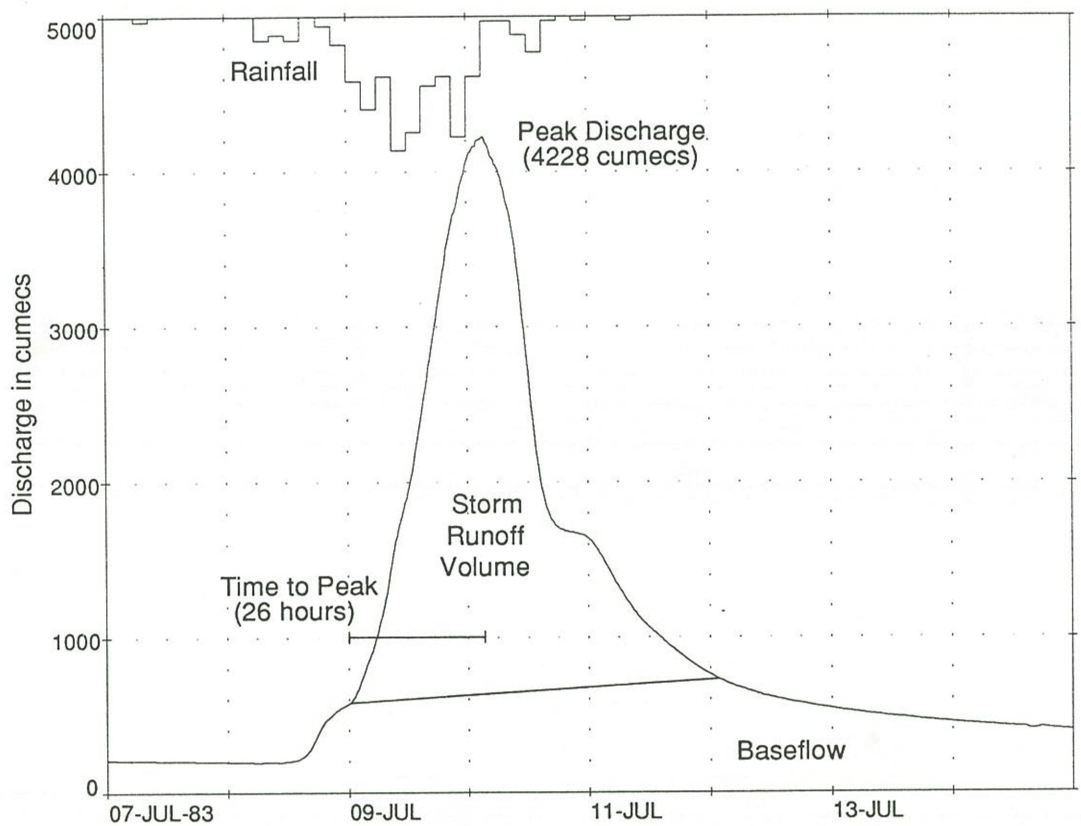


Figure 6.3 Hydrograph terms (hydrograph for the Grey River at Dobson, West Coast; Waipuna 3-hourly rainfall totals shown, maximum 17 mm).

used statistical index is the return period of a flood, which is related to the probability that a given discharge is equalled or exceeded. For example, a flood peak discharge with a 1% probability of being equalled or exceeded in any one year (a 1% annual exceedance probability, AEP) is often described as the flood with a 100-year return period. This is commonly called the 100-year flood.

It should not be assumed that a flood discharge with a 100-year return period will occur or be exceeded once every 100 years. In fact, the probability of this discharge being exceeded at least once in a given 100-year period is only 63%. On the other hand, there is a probability of 1.7% that two or more events which exceed the flood with a 100-year return period will occur in a given 20-year period. Hence, there is a finite probability that rare events may recur in a short time interval, as the occurrence of two major floods, considered to have return periods of 13 years and 36 years, in Greymouth during 1988 demonstrated.

The most common method of analysing flood frequencies for a given location is to fit a statistical distribution to a series of annual maximum flood peaks. Another common method, the "peaks-over-threshold" approach, considers the n largest flood peaks during a period which is usually between n/2 and 2n years long. Hence, the difference between the annual series and peaks-over-threshold methods is that the annual series uses the single largest flood each year, whereas the peaks-over-threshold may use none or several peaks from any one year. For large events, the two approaches give similar results.

Before frequencies are analysed, the annual series must be checked for data errors, using records from nearby catchments or rainfall data. For example, a period of missing record might have contained the largest flood for that year. If an accurate estimate cannot be made, then no other value should be used for that year.

A key assumption of flood frequency analysis is that the annual maximum flood series exhibits no trends or cycles, and that flood peaks are drawn randomly from the same statistical distribution that has applied in the past and will continue to apply in the future. McKerchar and Pearson (1989) and Withers and Pearson (1991) found no conclusive trends in recent New Zealand flood data, but growing evidence for the global warming indicates that the possibility of future trends should not be neglected.

Grant (1965, 1977, 1985) has found evidence for long-term variations in storminess, erosion, and flood flows, on time scales of decades and centuries. Our improving knowledge of El Nino-Southern Oscillation climate patterns is providing greater awareness of variations in the frequency and magnitude of floods and droughts, on a time scale of a few years. Recalculations of flood frequencies for a given river, as the length of available record increases, have shown that data collected during a given period may not be representative of longer periods. For example, the flood with a 50-year return period in the Waimakariri River was calculated in 1951 to be 3950 m^3/s, using data for the period 1931 to 1950. In 1971, using 20 years of additional data, the flood was recalculated to be 3900 m^3/s, and in 1991 it was recalculated again to be 3700 m^3/s.

Several statistical distributions may be fitted to the annual flood series in a river, as a basis for estimating the discharge of a flood with a given recurrence interval. The Extreme Value Type 1 distribution (EV1 or Gumbel distribution) is satisfactory for the annual flood series in most New Zealand rivers (McKerchar and Pearson, 1989, 1990). This distribution has been widely applied in New Zealand since its introduction by Benham (1950).

The EV1 distribution (Figure 6.4A) is not symmetrical, like the more familiar normal distribution, which is bell-shaped. The EV1 distribution is positively skewed, with most floods in the range of lower values, and fewer in the tail of higher values. Other statistical distributions used for flood frequency analysis have longer and/or thicker tails, reflecting differences in the flow regimes of the rivers to which they are applied.

EV1 and normal distributions (Figure 6.4A) can be converted into cumulative frequency distributions (Figure 6.4B). For convenience of analysis, such a curve may be converted into a straight line on a graph, using a special function of annual exceedance probability. More simply, the annual

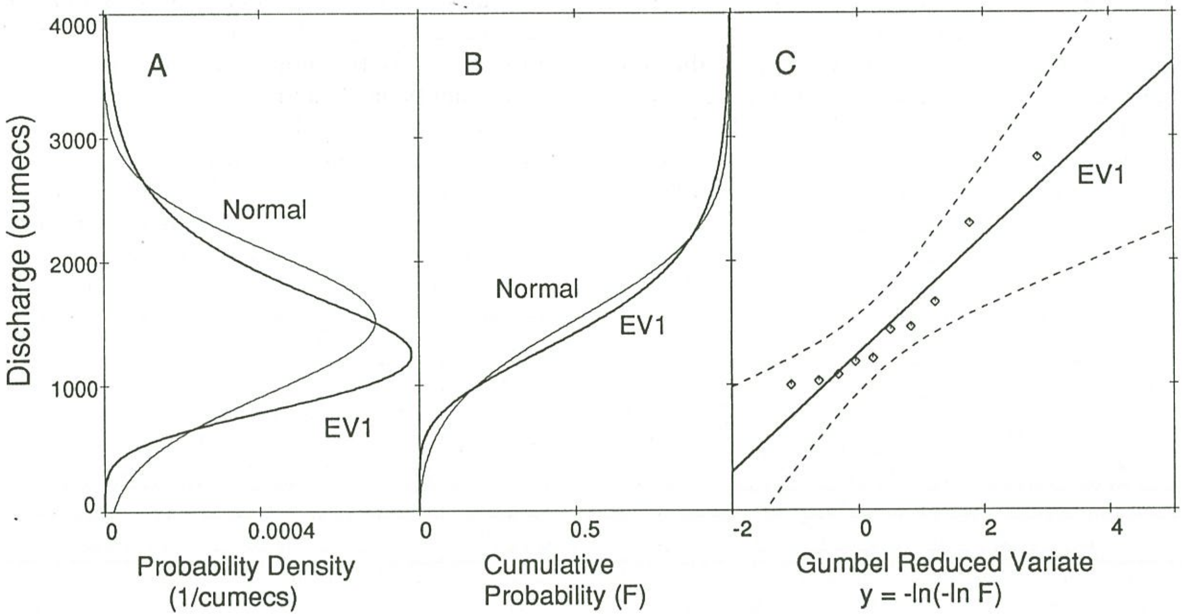


Figure 6.4 (A) EV1 and Normal distribution probability densities. (B) Cumulative versions (F) of 6.4A with arithmetic probability scale. (C) Gumbel probability plot of the Waimakariri River at Old Highway Bridge (Canterbury) annual maximum flood peaks, 1971-80, with 95% confidence limits.

maximum floods for a given river may be ranked in order of size and plotted on special EV1 probability paper (Gumbel paper), to see whether they lie on a straight line, and are therefore satisfactorily fitted by the EV1 distribution (Figure 6.4C).

The formula used most often to calculate plotting position is that of Gringorten (1963):

$$F(Q_i) = (i-0.44)/(n+0.12), \text{ and} \qquad (6.1)$$

$$y(Q_i) = -ln(-ln(F(Q_i))), \qquad (6.2)$$

where Q_i is the ith flood and n is the total number of floods. Hence, with n = 10 annual floods, the smallest is plotted at $F(Q_1)=0.055$ and $y(Q_1)=-1.063$, and so on (Table 6.1; Figure 6.4C).

A curve or line may be fitted to data graphically or mathematically. Methods of mathematically fitting the EV1 distribution include moments, least squares, probability-weighted moments, and maximum-likelihood. Moments methods insert sample moments such as the mean and standard deviation into algebraic equations that link moments and EV1 parameters to estimate the parameters. The maximum likelihood method selects the EV1 line which has highest probability of being correct. Each approach gives slightly different lines.

The equation for plotting an EV1 line is

$$Q = u + ay, \qquad (6.3)$$

where u and a are the parameters of the EV1 distribution and y is the Gumbel Reduced Variate, given by

$$y = -ln(-ln(1-AEP)). \qquad (6.4)$$

For example, for a flood with a 100-year return period, AEP=0.01, and y=4.6.

Year	Peak m^3/s	Rank (i)	$y(Q_i)$
1981	1460	7	0.836
1982	1440	6	0.513
1983	1660	8	1.232
1984	2830	10	2.866
1985	1080	3	-0.318
1986	1030	2	-0.626
1987	1180	4	-0.044
1988	2300	9	1.787
1989	1210	5	0.227
1990	1000	1	-1.063

Table 6.1 Annual maximum flood peak series (1981-90) for the Waimakariri River at the Old Highway Bridge (drainage basin area - 3210 km^2, site 66401, Walter 1990).

Since the true EV1 parameters are estimated using imperfect sample data, there is sampling error s_Q associated with an EV1 line. The values $Q \pm 2 * s_Q$ are the 95% confidence limits around the EV1 line, between which there is a 95% probability that the true EV1 line is located (Figure 6.4C). The confidence limits do not take account of systematic errors, for example, if the EV1 is the incorrect statistical distribution, or if there are errors in discharges estimated from an inaccurate rating curve.

The EV1 distribution belongs to a family of extreme value distributions, the Generalised Extreme Value (GEV) Distribution (Hosking et al., 1985). It has three parameters which describe its shape, the third one determining whether the distribution is of Type 1, 2, or 3 (EV1, EV2, or EV3). When the third parameter k has a value of zero the distribution is Type EV1 (that is, it becomes a two-parameter distribution). When $k < 0$ the distribution is termed EV2, and when $k > 0$ the distribution is termed EV3. On Gumbel paper an EV2 distribution curves upwards, and an EV3 distribution curves downwards. Hence, the flood data for the Waimakariri River (Figure 6.4C) tend to be better fitted by an EV2 distribution. A statistical test can be used to test whether the EV1 distribution is a better fit than the EV2 or EV3 alternatives (Hosking et al., 1985).

In contrast to other statistical distributions used for extremes, the EV1 distribution has theoretical justification, although strict application of extreme-value theory requires that the number of events (floods) in each time period (years) is large (so that the maximum for each year is the maximum of say 10 or more events). The distribution has been successfully used for extreme rainfalls in New Zealand (Tomlinson, 1980), and floods can be viewed as storm rainfalls minus catchment losses.

Three-parameter distributions such as the GEV should not be used to analyse flood frequency at single sites, unless the annual series is at least 30 years long, since sampling errors are much larger than for two-parameter distributions. Where the annual series is less than 30 years long and the EV1 distribution provides an unsatisfactory fit, regional methods which use information from neighbouring and similar catchments should be used. Indeed, frequency analysis of single sites should, if possible, be compared with the results of a regional analysis. Regional information is needed to satisfactorily estimate the probabilities of exceedance of extreme floods for sites with ten or fewer years of record.

Regional Estimates of Flood Frequency

Beable and McKerchar (1982) divided New Zealand into several regions with comparable flood frequencies, using the method adopted for the UK Flood Study (Natural Environment Research Council, 1975). Regression equations were developed for each region, to predict mean annual flood ($\overline{Q}$), using as independent variables such properties as catchment area (A) and the 24-hour rainfall with a 2-year return period (I). For example, the best estimate of mean annual flood for the west coast of the South Island is given by:

$$\overline{Q} = 0.0233\, A^{0.94}\, I^{0.99} \qquad (6.5)$$

For example, an ungauged tributary of the Buller River, the Ohikanui River, draining the northwestern Paparoa Range, with an area of 125 km^2 and a 24-hour rainfall (2-year return period) of 240 mm, has a mean annual flood estimated as 495 m^3/s.

Dimensionless factors were also estimated for each region, which by multiplying by the mean annual flood provide estimates of floods with other return periods. For the west coast of the South Island, for example, the factor for a flood with a return period of 10 years (Q_{10}) is 1.41. So the regional estimate for the Ohikanui River is Q_{10} = 1.41 * 495 = 700 m^3/s.

McKerchar and Pearson (1989, 1990) reviewed New Zealand flood frequencies, using nearly ten more years of data than were available to Beable and McKerchar. For mean annual flood, 343 annual series were used, each of which had at least 6 values. Mean annual flood is a partial function of catchment area (Figure 6.5), with the least squares regression equation given by:

$$\overline{Q} = 2.04\,A^{\,0.808} \qquad (6.6)$$

However, for a given catchment area mean annual flood varies widely, for example over two orders of magnitude for a catchment area of 1,000 km^2.

Maps of a form of specific discharge, $\overline{Q}/A^{0.8}$, plotted for the 343 catchments indicated smooth trends over New Zealand, for which contours

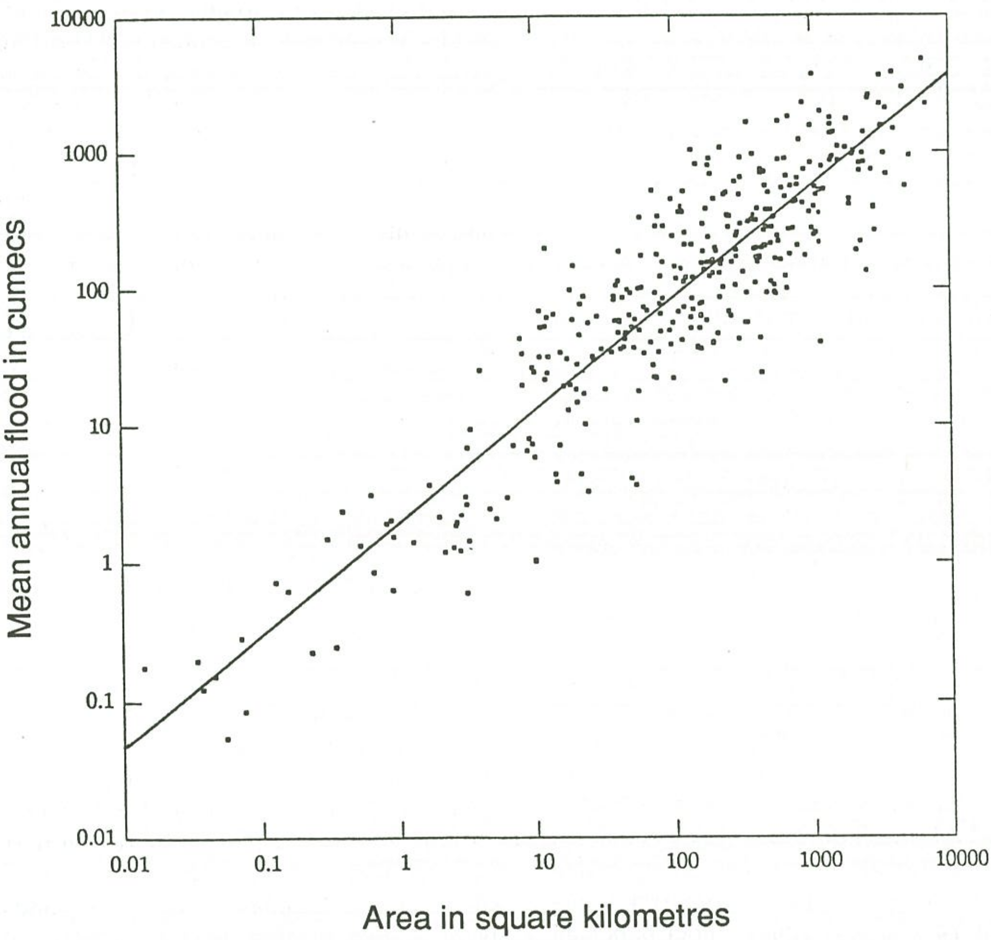


Figure 6.5 Mean annual flood (Q, m^3/s) plotted against drainage basin area (A, km^2) for 343 New Zealand water-level recording sites (from McKerchar and Pearson, 1989).

could be drawn (Figure 6.6). (If flood frequency regions truly existed, they would be indicated by areas having relatively constant values of $\overline{Q}/A^{0.8}$, rather than smooth trends.) The contours generally reflect the pattern of annual rainfall (NZ Meteorological Service, 1985) and rainfall intensity (Tomlinson, 1980), with high values along the main mountain ranges and around the North Island volcanoes, and low values in areas of rain shadow. Low values in the central North Island, north of Taupo, are attributed to the layer of absorbent volcanic ash in that area.

Use of the contour maps in Figure 6.6 to estimate mean annual flood has a precision of ± 22% for 95% of the catchments used. For example, the estimate of mean annual flood for Waipa River at Whatawhata, which drains 2,820 km^2 of the King Country, is $0.6*(2820)^{0.8} = 345\ m^3/s$ (± 22%). This compares favourably with mean annual flood of 388 m^3/s derived from 18 years of flow measurements.

For 19 sites, map-derived estimates of mean annual flood were 70% greater than the estimates derived from flow measurements; 13 of the sites have catchment areas less than 100 km^2. However, a subsequent study of more than 100 catchments with areas less than 100 km^2 (McKerchar, in press) could not improve regional estimation of mean annual flood beyond that of the map approach.

Although the map approach provides the best available estimates of mean annual flood for New Zealand catchments, multiplicative regression equations, as used by Beable and McKerchar (1982), are still valuable. They can be used to analyse the spatial and temporal variations in mean annual flood estimates, to seek improvements in the locations of the recording stations in a region (Pearson, in press).

It is possible to estimate the discharge of floods with other return periods, using estimates of mean annual flood obtained from contour maps (Figure 6.6). The EV1 distribution is satisfactory for 228 of the 275 catchments used by McKerchar and Pearson (1989, 1990), and values of $Q_{100}/\overline{Q}$, (where Q_{100} is the 100-year return period flood) representing the slope of their EV1 distributions, have been plotted on maps of New Zealand (Figure 6.7).

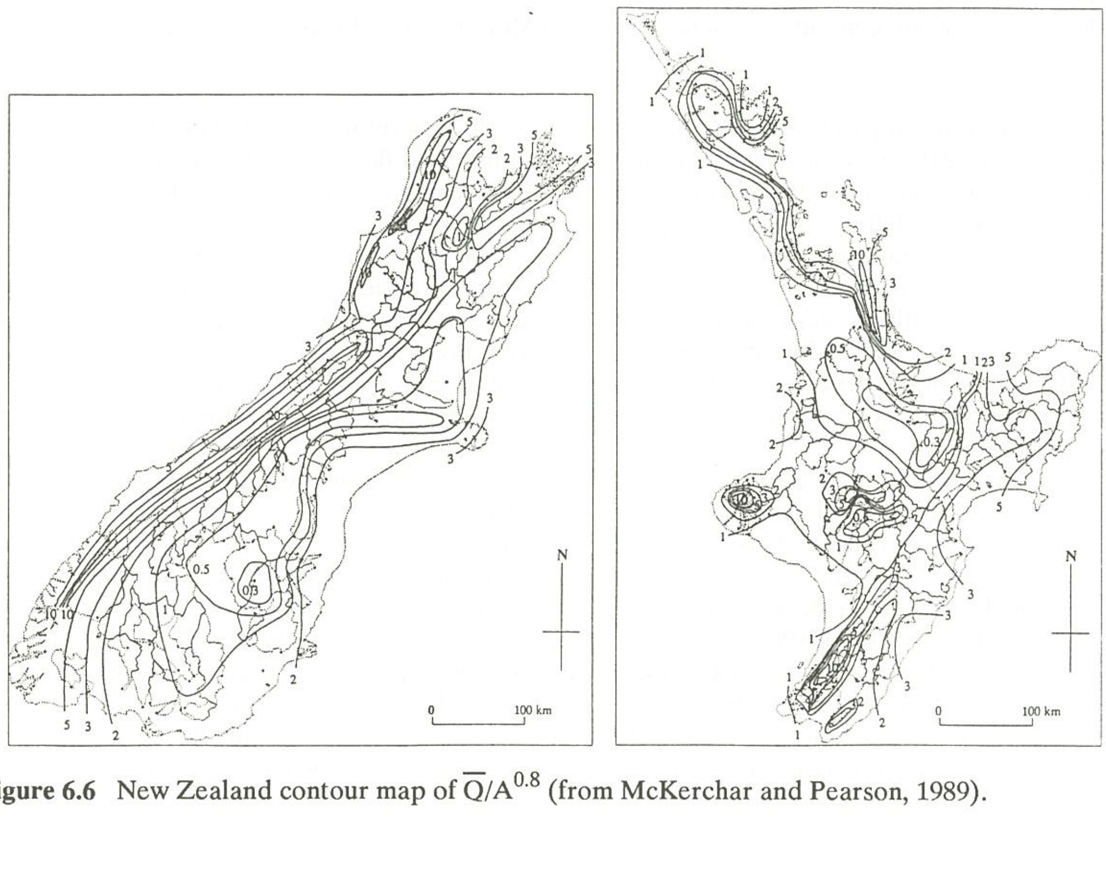


Figure 6.6 New Zealand contour map of $\overline{Q}/A^{0.8}$ (from McKerchar and Pearson, 1989).

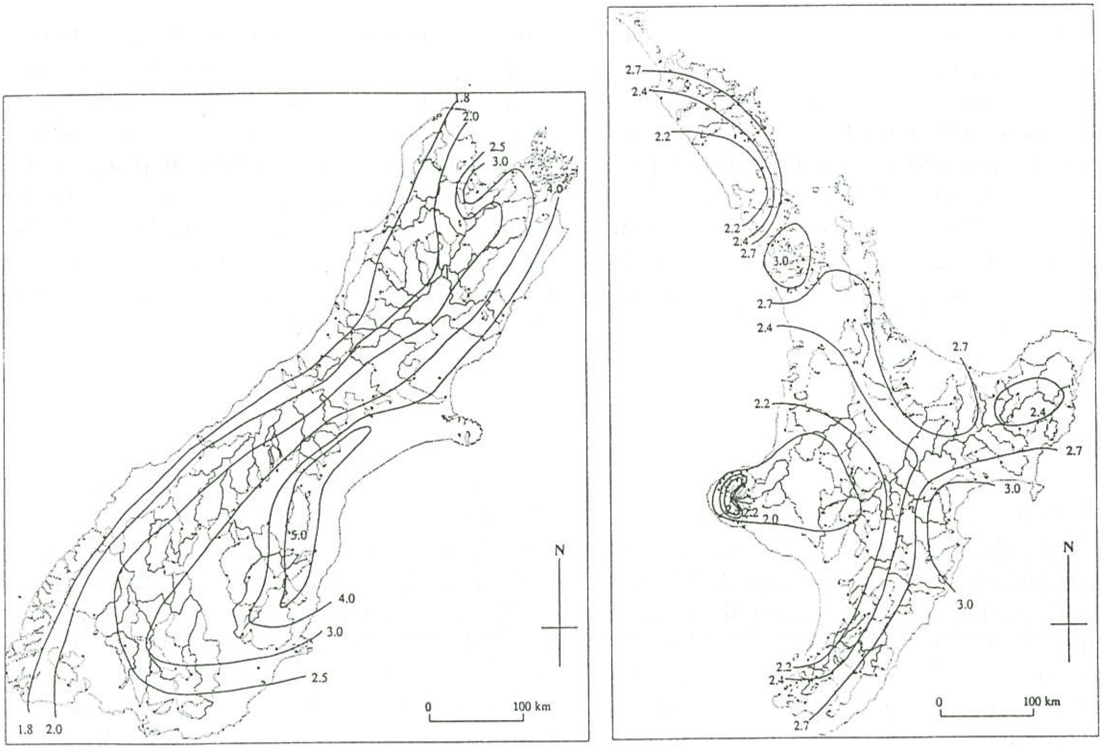


Figure 6.7 New Zealand contour map of $Q_{100}/\overline{Q}$ (from McKerchar and Pearson, 1989).

Smooth variations are apparent, as with $\overline{Q}/A^{0.8}$. Values are low along the western sides of the main mountain ranges and the North Island volcanoes, and are high where rainfall is low and infrequent, for example in South Canterbury. This reflects the fact that year-to-year variations in the size of the largest floods are relatively small in wet areas, which receive a regular succession of heavy rainfalls. On the other hand, in dry areas rainfall is less frequent, but occasionally comes in intense frontal or convectional storms.

The range of variation of the 100-year return period flood to the mean annual flood is much less in the North Island (2 to 3) than in the South Island (1.8 to >5). Values increase from west to east across the South Island and, to a lesser extent, from southwest to northeast in the North Island.

Use of the maps in Figure 6.7 to estimate $Q_{100}/\overline{Q}$ has a precision of ± 17%. For the Waipa River at Whatawhata, the map estimate is 2.3, whereas the estimate using 18 years of discharge measurements is 1.98. Therefore an estimate of the 100-year return period flood using the maps is 2.3*345 = 793 m^3/s , compared with 768 m^3/s from at-site EV1 analysis.

For return periods other than 100 years, the assumption of the two-parameter EV1 distribution allows mean annual flood and 100-year return period flood to be connected by a straight line on Gumbel paper, which enables estimates to be made of the flood discharge for any desired return period T (Figure 6.8). Alternatively, the formulae given in McKerchar and Pearson (1989, 1990) may be used.

McKerchar and Pearson's study assumed that annual maximum flood series in New Zealand are fitted by the EV1 distribution. In fact, a comparison of the 275 actual flood series with 275 randomly generated EV1 series indicated that the real floods tend to be better fitted by an EV2 distribution with a value of the parameter k of -0.07, which gives an upward curve on Gumbel

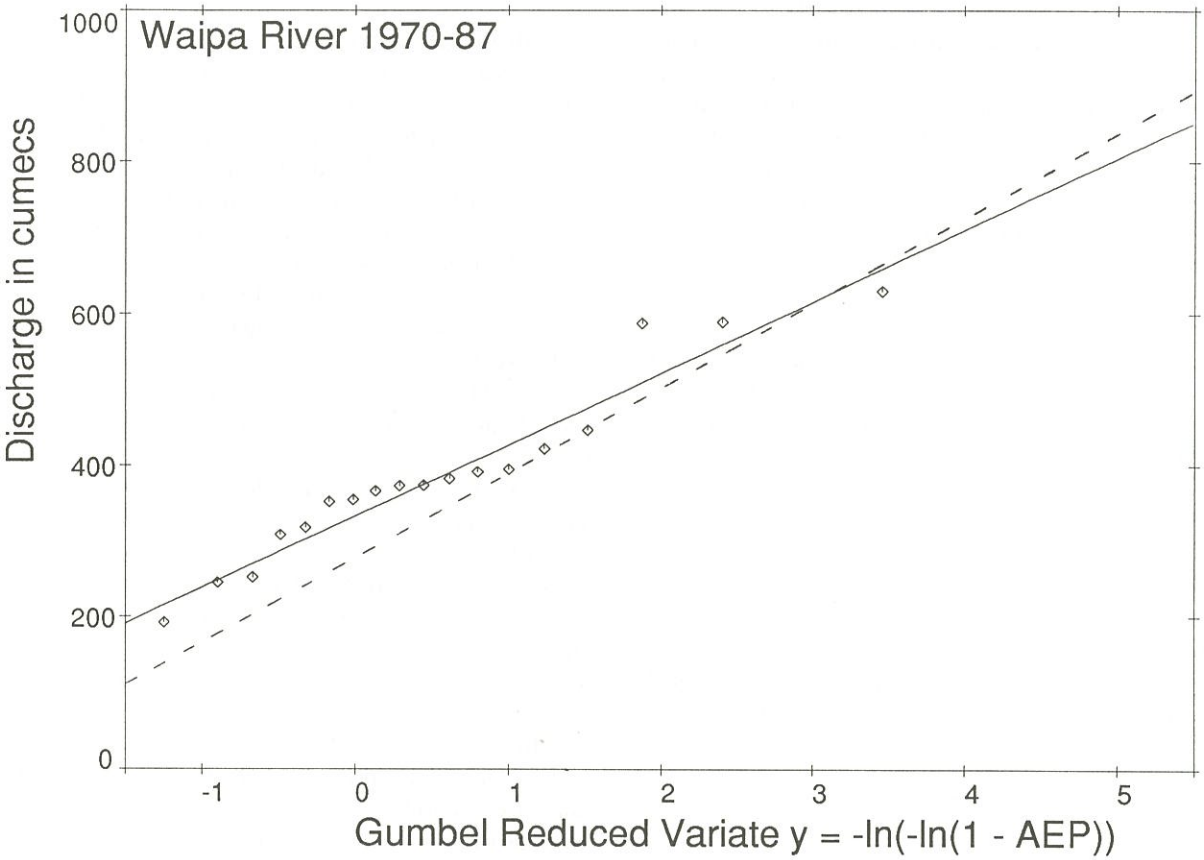


Figure 6.8 Annual maximum flood peaks (1970-87) for the Waipa River at Whatawhata, central west North Island, with at-site (full) and regional (dashed) EV1 lines.

paper. However, for return periods of 100 years or less, the difference between EV1 and EV2 flood estimates are negligible, with k = -0.07.

Risk Analysis

If the design life of a proposed structure such as a stopbank is known, and the acceptable risk that it might fail or be damaged can be specified, it is necessary to estimate the return period of the flood which the structure must be able to withstand - the design flood. The binomial risk formula may be used for this purpose. It enables the engineer to estimate the risk or probability that a flood with a specified return period will be equalled or exceeded during a specified interval, such as the next 200 years. The risk r of the T-year return period flood being exceeded at least once in the next L years is

$$r = 1 - (1-1/T)^{L} = 1 - (1 - AEP)^{L} \qquad (6.8)$$

For example, the probabilities that the flood with a 100-year return period will be equalled or exceeded at least once in the next 20, 50, 100, and 200 years are 18%, 40%, 63%, and 87% respectively. Hence, for instance, if the design life of a stopbank is 100 years and the acceptable risk of failure during that period is 1%, the return period of the design flood is 9,950 years; that is, the structure must be designed to withstand the 9,950 year flood. Such a low level of risk might be justified when, for example, a major city like Christchurch is threatened by overtopping and failure of a stopbank.

One of the main assumptions of flood frequency analysis is that conditions in which data were collected will continue into the future. Climate change, land-use change, and other factors may invalidate this assumption, and the size of the flood with a given return period may change with time.

The effects of 5%, 10%, 15%, and 20% increases in flood peaks for the Hutt River at Birchville (catchment area of 427 km^2) over various periods have been examined (Table 6.2). For example, a linear increase in flood peak magnitude resulting in a cumulative 20% change in 200 years time: the risk of exceedance during the next 200 years of the present 100-year and 1000-year return period floods (1,760 m^3/s and 2,360 m^3/s respectively) would increase from 87% to 98% and from 18% to 39%, respectively.

Rainfall-runoff Models

Where there are long rainfall records but short discharge records, mathematical rainfall-runoff models (Chapter 7) may be used for estimating the size of design floods. They provide an estimate of the full design hydrograph, rather than just the peak discharge. They can, therefore, be used to estimate the durations and volumes of high flows, which might be important, for instance, to the river control engineer who must design a stopbank to withstand the erosion by prolonged high flow.

On the other hand, rainfall-runoff models require data on, or estimates of rainfall, and many assumptions are made in calibrating a model for a catchment and in extrapolating to large events with high return periods. Model parameters need to be identified for a catchment during calibration, using rainfall and runoff data from at least one recorded flood. The set of calibration parameters can then be used to predict design flood flows under an assumed set of extreme conditions, such as heavy rainfall onto an already saturated catchment.

A flood with a particular return period can be caused by a range of combinations of rainfall and antecedent catchment wetness. A short burst of intense rainfall can produce a sharply peaked flood with a maximum discharge as great as that reached in a flood which results from less intense rainfall with a longer duration. Similarly, heavy rainfall onto a dry catchment can cause a flood peak as large as that caused by less intense rainfall onto a saturated catchment.

As an example, a rainfall-runoff model has been calibrated for the Hutt River catchment, using data from the Birchville gauging station (Figure 6.9). A design rainfall with a duration of 6 hours was used as input to the model, to obtain design hydrographs at different points along the river (Figure 6.10). The rainfall chosen was sufficient to produce at Birchville the flood with a return period of 100 years, which has been estimated by flood frequency analysis at 1,760 m^3/s.

		Flood magnitude increase scenario:				
		0%	5%	10%	15%	20%
(a)	1760 cumecs L (years)					
	50	40	45	51	57	63
	100	63	70	76	81	86
	200	87	91	94	96	98
	500	99	100	100	100	100
(b)	2360 cumecs L (Years)					
	50	4.8	6.1	7.6	9.4	12
	100	9.5	12	15	18	22
	200	18	22	27	32	39
	500	39	46	54	62	70

Table 6.2 Risks (%) of exceedance of flows of 1760 and 2360 cumecs (present Q_{100} and Q_{1000}) for the Hutt River at Birchville, for scenarios of a 0%, 5%, 10%, 15% and 20% increase in flood magnitude over various design life years (L).

Maximum Floods

Estimation of the most extreme flood possible is required for designing structures whose flooding or failure would have disastrous consequences, such as nuclear power stations or large hydro-electric dams. Several methods are used to estimate the largest flood that a catchment can dis-

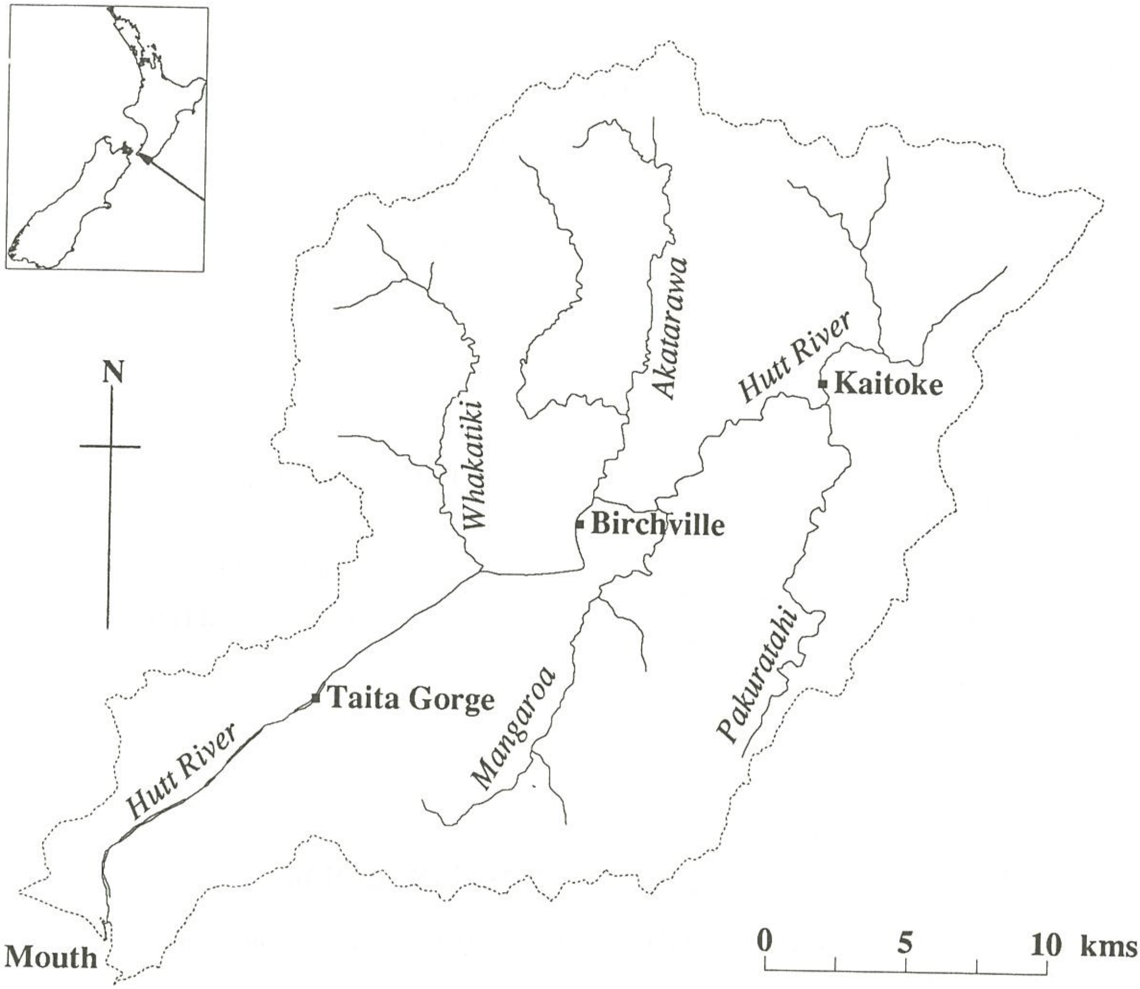


Figure 6.9 Drainage basin of the Hutt River, Wellington, showing various tributaries to the main channel and location of the Birchville water-level recorder.

charge in response to the heaviest possible rainfalls.

The maximum rainfall is termed the probable maximum precipitation and the maximum flood is the probable maximum flood. However the methods normally used to estimate probable maximum precipitation and probable maximum flood are based on physical rather than statistical considerations, and therefore take little account of the probabilities of occurrence.

Methods to estimate probable maximum flood normally use a rainfall-runoff model for a catchment, together with estimates of probable maximum precipitation (Chapter 4). Because of the assumptions necessary in the estimation of probable maximum precipitation and rainfall- runoff modelling, the accuracy and precision of estimates of probable maximum flood vary from method to method, but cannot be precisely quantified. A simple method of estimating the probable maximum flood in a catchment is to look at the historical occurrence of large floods. Figure 6.11 is a plot of the greatest discharge ever recorded in 343 New Zealand catchments, as a function of drainage area. An envelope curve has been drawn through

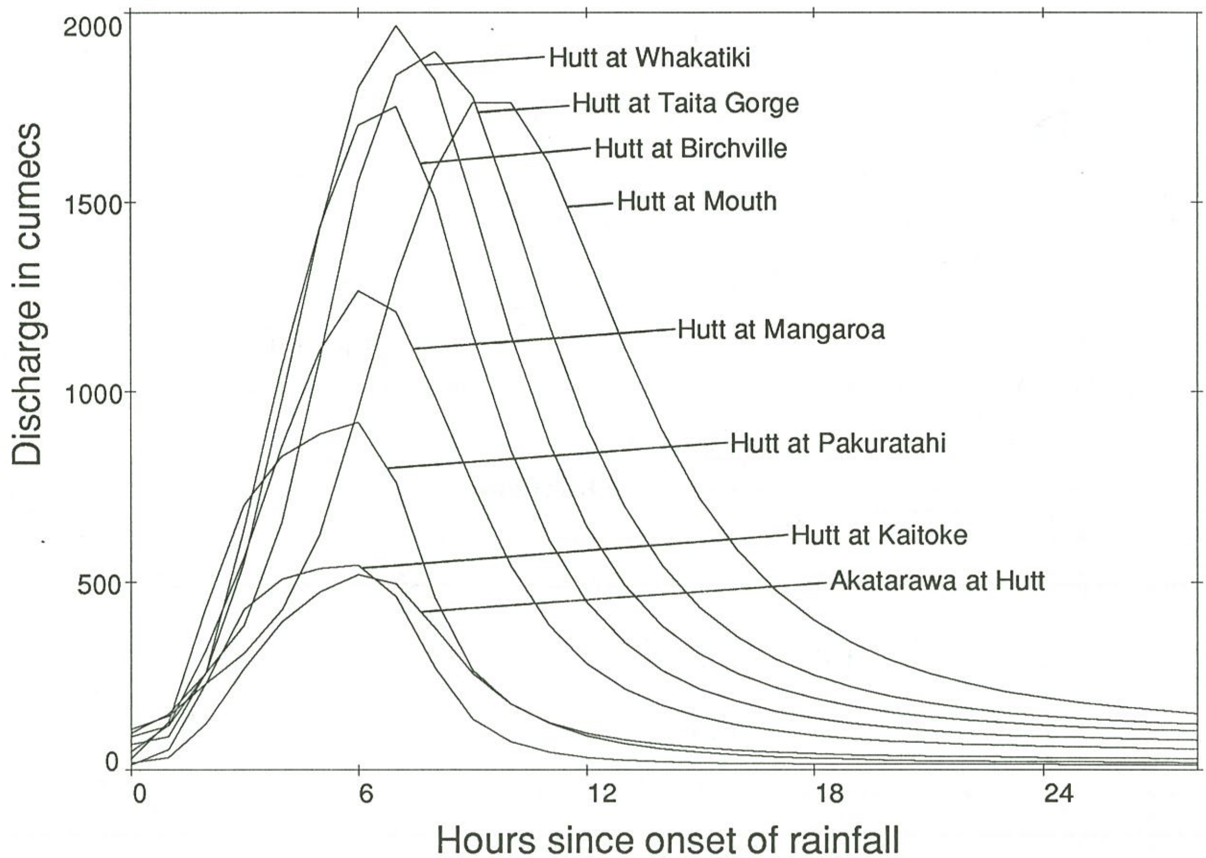


Figure 6.10 100-year flood hydrographs for the Hutt River, estimated from a rainfall-runoff model using 6-hour duration rainfalls.

the three most extreme floods in the plot. Discharges for the Haast River and Cropp River, which flow from the western Southern Alps, resulted from heavy rainfall during very strong northwesterly air flows, and may be approaching the probable maximum flood. However, Costa (1987) reported that measured maximum peak floods were on average 75% of estimated probable maximum floods for nine catchments in the United States which had flood maxima near his upper envelope curve.

Flood Forecasting

Flood forecasting is required to warn people with homes, assets and livestock in flood-prone areas to evacuate and to implement emergency contingency plans. Most flood protection structures in New Zealand have finite probabilities of overtopping or failure, so that forecasts of impending floods are a necessity even in locations where flood protection schemes have been provided.

The question "When will the next flood occur and how large will it be?" can only be answered in the hours or days before a flood, by examining synoptic weather conditions and monitoring rainfall and river flows. In New Zealand, most flood forecasting is carried out by hydrologists employed by regional councils. They use telemetered rainfall and water-level recorders to obtain information on present conditions in their catchments, in conjunc-

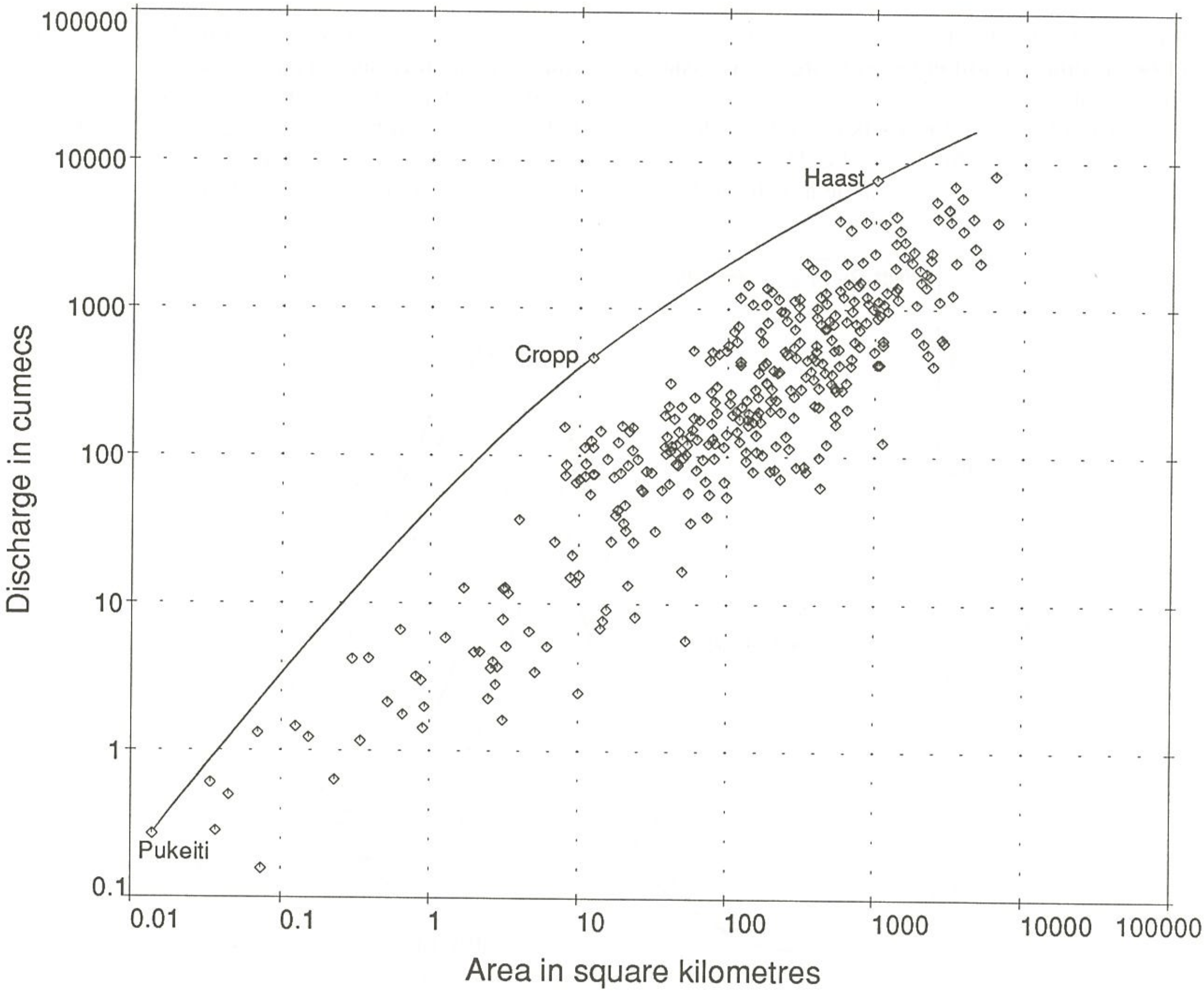


Figure 6.11 Maximum recorded flood peaks at 343 New Zealand drainage basins (taken from data in McKerchar and Pearson, 1989).

tion with weather forecasts provided by the Meteorological Service of New Zealand.

Estimates of the timing and size of flood peaks often rely heavily on the past behaviour of a river. Objective, automated methods use mathematical models to make predictions on the basis of telemetered information on rainfall and water levels. Such objective methods used in New Zealand include the linear systems flow-to-flow approach for the Grey River (Goring, 1984), conceptual catchment storage-rainfall-runoff models for the Waipa River (Waikato), multiple linear regression of flood peaks on moving-average rainfall data for the Wairau River (Rae and Wadsworth, 1990), and empirical rainfall-runoff routing for the Waimakariri River (Griffiths et al., 1989).

A survey of flood forecasts (Pearson and Jordan, 1991) showed that objective forecasting methods were more reliable than subjective methods, in that objective forecasts made more than three hours before a flood peak were three times more likely to be within 20% of the actual peak than subjective forecasts. Pearson and Jordan showed the level of reliability for objective forecasts to be about 80%, using the Ibbitt et al (1990) rainfall-runoff model for three North Island catchments.

Figure 6.12 shows forecasts made for a flood in one of these catchments, the Hutt River at Birchville. Forecasts were made hourly, and were based on the assumption that rainfall ceased at the time that the forecast was made, which was clearly incorrect. The forecasts two hours and one hour before the actual peak provided a usable estimate of peak flow but an inaccurate estimate of hydrograph shape, because of the incorrect as-

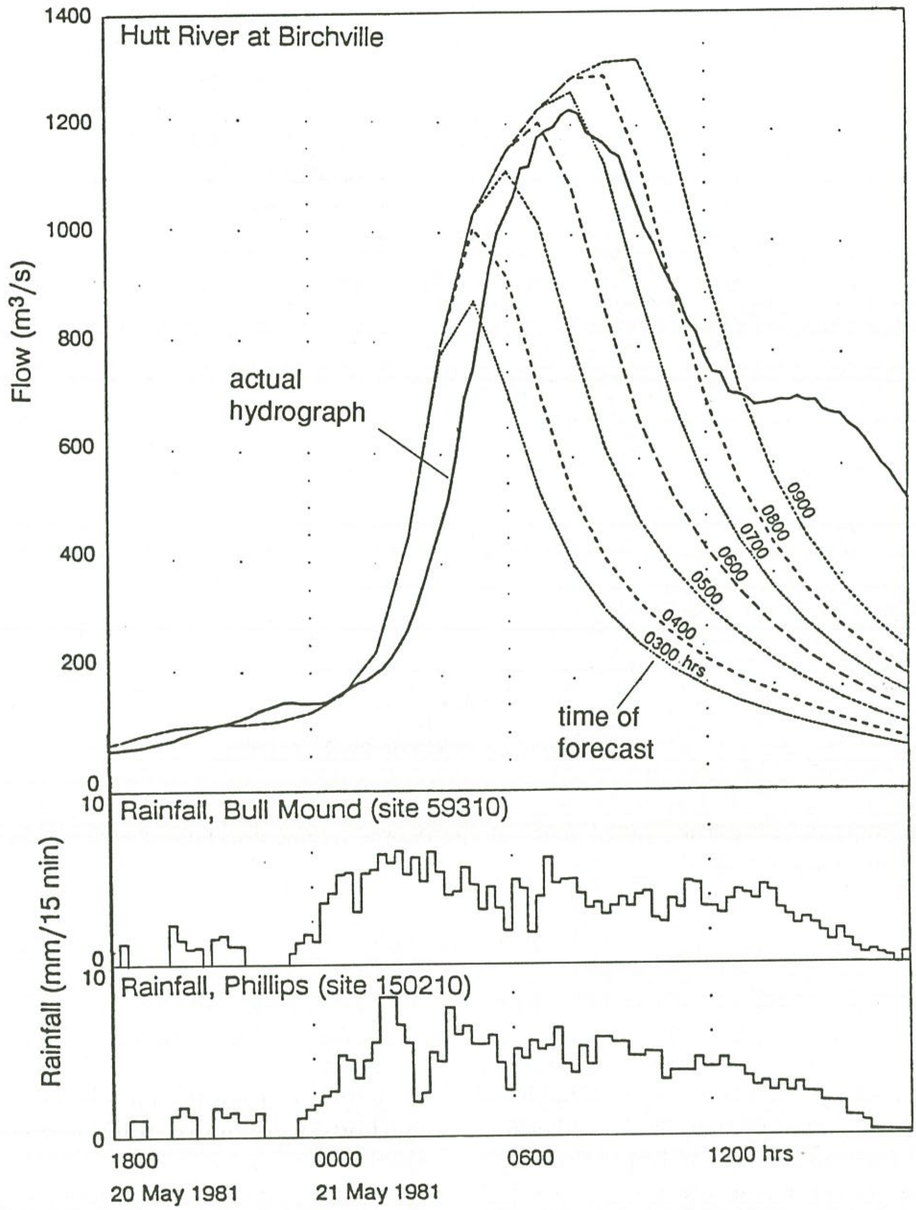


Figure 6.12 Flood forecasts for the Hutt River at Birchville, Wellington, from a rainfall-runoff model made at hourly intervals after the onset of heavy rain.

sumption about rainfall. However, forecasts made at and one hour after the peak were inaccurate, in that they predicted peaks higher and later than the actual peak. Flow forecasts can be markedly improved by including rainfall forecasts (Bertoni et al., 1992).

Flood forecasts obtained by rainfall-runoff modelling are less accurate than those obtained by applying a hydrograph observed at an upstream monitoring site to sites downstream. This is because of difficulties in modelling variables such as subsurface flow. However, forecasting methods based upon prediction of runoff from rainfall are needed because they provide earlier warning. In headwater catchments there is no alternative, since flow-monitoring stations do not exist further upstream.

Telemetered raingauges and weather radar show where rain is falling in a catchment. This greatly assists the analyst to predict flows through a channel network, and to interpret the results of objective forecasting methods. In other countries methods are being developed to combine quantitative precipitation forecasts, satellite imagery of storm systems, weather radar, and telemetered rainfall and streamflow to provide flow forecasts. This may happen in due course in New Zealand, although this country is hampered by the lack of meteorological observations for the surrounding oceans.

Low Flows

In low flow hydrology, hydrologists are principally interested in whether or not a river or stream can supply a given demand for water. Water may be needed for domestic or industrial use, for irrigation of farmland, for hydroelectric power generation, for recreational river use, or to maintain wildlife habitat. If demand cannot always be met, the amount of water which must be stored in order to meet probable water deficits must be estimated. To do this, frequency analysis of low flows and of water-deficiency volumes is required.

The index of low flow commonly used in frequency analyses is the annual minimum n-day mean flow, where n is often taken to be 7 days.

The form of the probability distribution of annual minimum flows is restricted, as flow minima cannot be less than zero, and annual series of flow minima are most often positively skewed. Distributions recommended for low flows include the Weibull (Nathan and McMahon, 1990) and the log-Normal distributions. Another possibility is the EV1 distribution, even though it has no lower bound. This may not be a serious problem, as the probability of values being less than zero in the EV1 distribution is usually much less than 0.1%.

The return period T of flow minima is taken as the reciprocal of the probability of non-exceedance:

$$T = 1/F = 1/(1\text{-}AEP) \qquad (6.9)$$

Application of the same procedure used for flood flows to annual minimum flows gives an estimate of q_T, the low flow with a probability of occurrence of 1/T. For the 100-year return period low flow (q_{100}), probability of exceedance in any one year is AEP = 0.99, and so the corresponding Gumbel Reduced Variate (Equation 6.4) is y = -1.527. For a specified design probability 1/T that water demand cannot be met, the value of q_T can be compared to the demand flow. If the value of a given low flow is large in comparison with the demand flow, then the river can meet the demand satisfactorily. However, if q_T is less than or approximately equal to the demand flow, then the river cannot be expected to meet demand without some form of flow regulation or storage.

As an illustration of frequency analysis of low flows, Table 6.3 lists annual minimum 7-day mean flows for the Hutt River at Birchville, for the period 1971-90. These values are plotted on Gumbel paper in Figure 6.13; the EV1 distribution provides an acceptable fit.

The 100-year return period, minimum 7-day mean flow q_{100} is estimated to be 840 l/s. Estimates of minimum 7-day mean flows with return periods of 5, 10, and 20 years are 1,607, 1,346, and 1,155 l/s respectively. These do not differ greatly from each other, and are small in comparison to the mean flow of the river, which is 22,300 l/s. They vary from 7% to 5% of mean flow, which is a narrow range of variation in comparison with that

Year	q(l/s)	Year	q(l/s)
1971	1422	1981	2228
1972	2308	1982	2222
1973	1505	1983	1623
1974	2083	1984	2245
1975	2191	1985	1513
1976	3591	1986	2678
1977	3798	1987	1983
1978	1190	1988	2237
1979	3204	1989	1844
1980	5376	1990	2232

Table 6.3 Annual minimum 7-day mean flows for the Hutt River at Birchville (29818).

of design floods, which may range from 50 to 200 times the mean flow. For the Hutt River, a demand flow which equals only a small fraction of the mean flow can be satisfied without regulation. However, demand was larger - say, two thirds of mean flow - flows would decline regularly below the demand, leaving periods of water deficiency (Figure 6.14). The area between the demand flow and the hydrograph can be integrated to calculate water-deficiency volumes or "drought volumes", which would need to be supplied from storage to sustain demand flow.

Storage reservoirs are designed to meet the demand flow for an acceptable fraction of the time, consistent with the cost of construction. Reservoir capacity can be made equal to the water deficiency volume which is exceeded with an acceptable probability of occurrence or return period T. If the

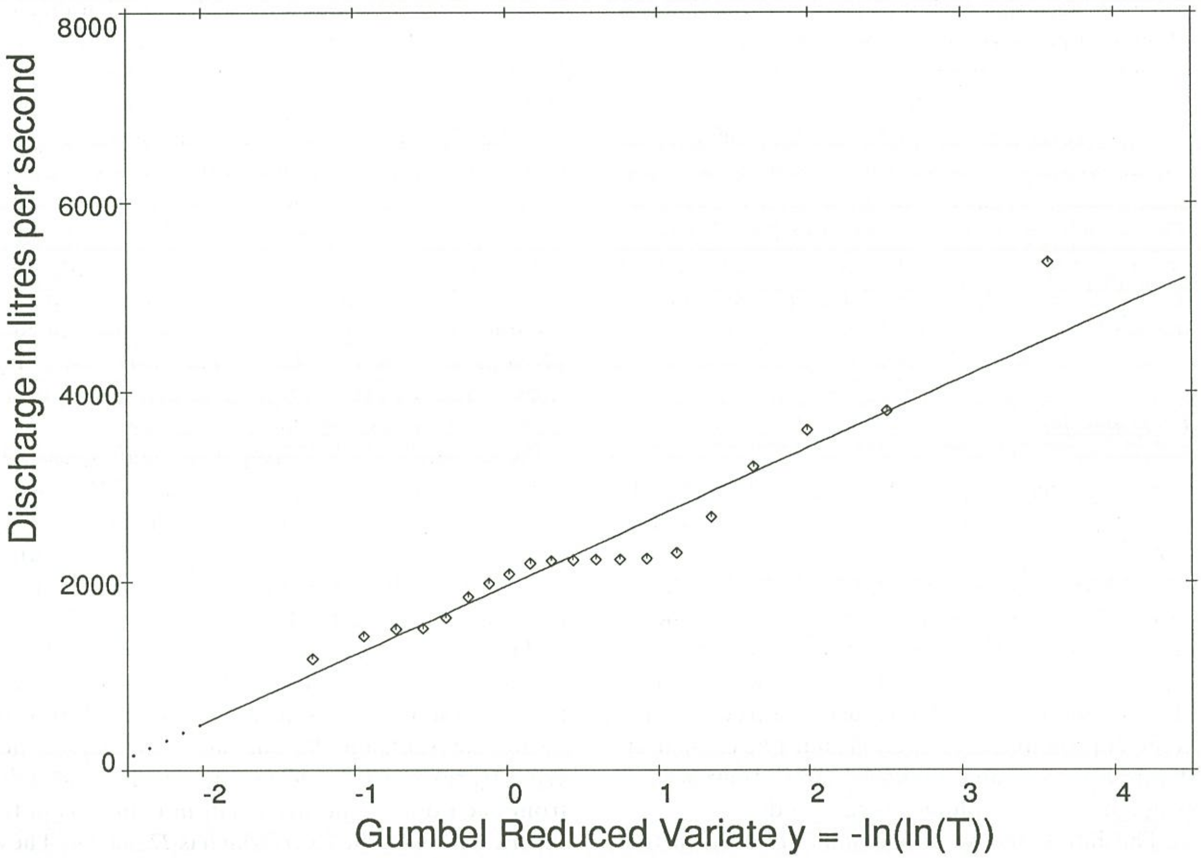


Figure 6.13 Gumbel probability plot of annual minimum 7-day mean flows for the Hutt River at Birchville, Wellington, (1971-90) with at-site EV1 distribution line.

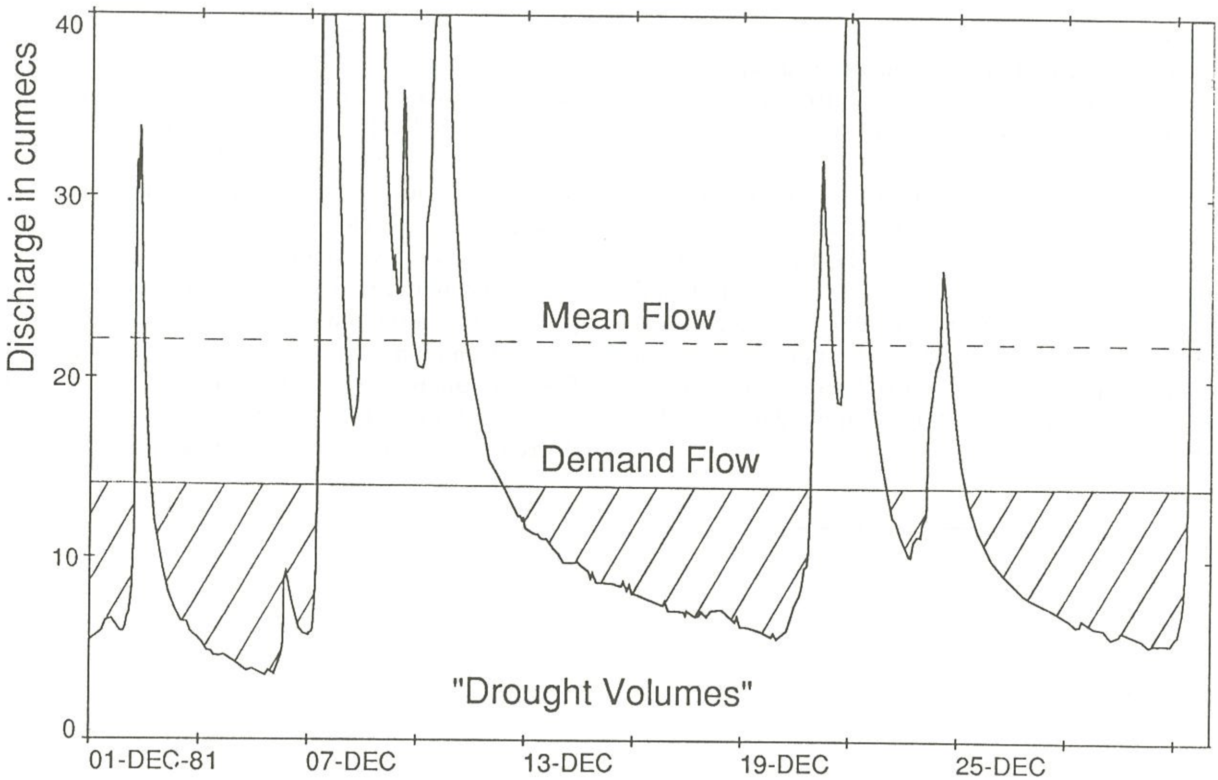


Figure 6.14 Diagram of drought volumes (hatched) for Hutt River at Birchville (December 1981) assuming water demand is two-thirds of mean flow.

storage provided is less than the actual water-deficiency volume in a particular year, then failure occurs. Hence, the annual maximum series of drought volumes can be analysed in the same way as for floods.

Such analyses assume constant demand. However, if there is a danger that the water-deficiency volume may exceed storage, then demand can be reduced, for example, by increased reliance on thermal rather than hydroelectric electricity generation.

As with flood flows, regional analyses of low flow can be used to predict flows for locations for which no flow measurements are available. In such a regional study, the low flow series at all sites in the region are analysed, and the minimum flows having some specified return period are estimated. These values, commonly in the form of specific low flows (discharge per unit area), can then be applied to catchments for which measurements are not available.

Two methods of predicting flow in ungauged catchments have been used in New Zealand. Flow may be measured in as many streams in an area as possible, during a period of low stable flow. Correlations may then be developed between the flows in streams which have continuous records of flow, and those which do not. For example, in a study of low flows of Taranaki rivers, McKerchar and Dymond (1981) plot six measured flows from the ungauged Waiaua River against corresponding flows at the same time from the gauged Punehu River (site number 36001, area 29.5 km^2), both rivers draining the southwestern side of Mt Taranaki. Using the resulting relationship with Punehu's 7-day q_5 estimate of 230 l/s gives a 7-day q_5 estimate of 800 l/s for Waiaua River at Wiremu Rd (drainage area of about 50 km^2).

A second method involves development of regression equations for low flows, using as independent variables catchment attributes such as catchment area, mean rainfall, hydrogeology, and slope. Hutchinson (1990) identifies 11 New Zealand low flow regions for which he provides regression equations for 7-day 5-year return period specific low flow. For the Taranaki region, the independent variables are annual rainfall and proportion of the basin with slope of 3 degrees or less. For Waiaua River at Wiremu Rd, using an annual rainfall of 2400 mm and slope proportion of 20%, gives 8.5 $l/s/km^2$ for specific 7-day q_5, or 425 l/s for 7-day q_5.

Low Flow Forecasting

Statistical analysis of low flows can provide estimates of the likely number of droughts of a given severity in a specified period (Griffiths, 1990), as well as other information for design purposes. An even more useful capability would be to forecast when a drought will actually occur, so that water use can be moderated and contingency plans set in place well beforehand.

By examining hydrographs of a river's past low flows, the analyst can construct a "master recession curve", which estimates how river levels will decrease over extended periods without rainfall.

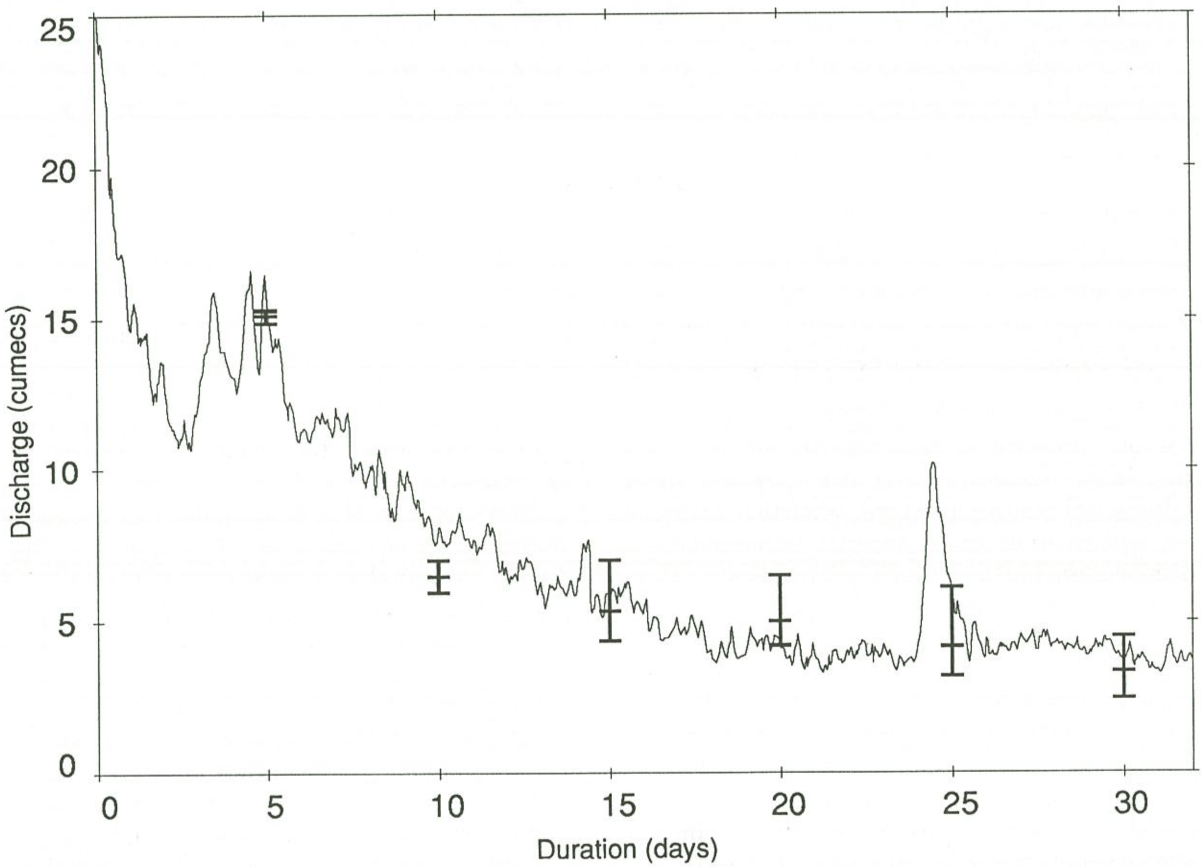


Figure 6.15 Low flow forecasts from a rainfall-runoff model (bars), made on 28 November 1986, and recorded flows (solid line), for the Ashley River at Gorge (site number 66204, drainage area 472 km^2). Each bar indicates minimum, mean, and maximum forecasts corresponding to different time distributions of the rainfall total (Shankar and Pearson, 1992).

The master recession curve can then be used in times of low flow to forecast expected river flows if no rainfall occurs. The most common mathematical model for recessions is an exponential decay of flow with time. Waugh (1970) shows that this model fits most Northland streams and rivers for periods of up to 10 days or so.

An alternative method for forecasting low flow recessions, particularly for longer periods, uses mathematical rainfall-runoff models. The advantage of this method is that different rainfall scenarios can be considered. For example, Figure 6.15 shows low flow forecasts for the Ashley River (Canterbury) for up to 30 days ahead, derived from a rainfall-runoff model using seasonal evaporation and 20% of mean rainfall totals for each forecast period (Shankar and Pearson, 1992).

Our growing understanding of the climate patterns associated with the El Nino-Southern Oscillation (Chapter 4) provides a potentially powerful tool for medium to long term forecasting of droughts and low flows. Several recent dry periods are ENSO-related. The Southern Alps drought of 1992 was associated with El Nino, a low Southern Oscillation Index. There were few weather systems bringing moist westerly winds to the west coast of the South Island (of national importance for inflows to hydroelectric lakes). The Canterbury drought of 1987-1989, on the other hand, was associated with La Nina, a high Southern Oscillation Index. Over a 14 month period there was a shortage of southerly weather systems that bring wet conditions to the east coast of the South Island, and an abundance of drying westerly winds (McGann, 1991).

Future Work

For floods, there is already good information about regional flood frequencies for return periods up to around 100 years. Estimation of longer return period floods remains difficult, because only relatively short records of river flows (generally less than 25 years) are available. Further work is still required to extend knowledge of the statistical attributes of floods, using physical approaches based on rainfall physics and catchment processes, and probabilistic models. To provide greater warning of impending floods, there is a need to link weather radar with information from telemetered raingauges and water-level recorders, to use in flood forecasting based on rainfall-runoff modelling.

Despite a history of crippling droughts in New Zealand, especially in the eastern parts of both islands, very little is known about drought occurrence and properties. The economic impacts of droughts often exceed those of floods, although they are more insidious and less dramatic. With global warming a reality in New Zealand (whether or not induced by the greenhouse effect), droughts may become more severe and more frequent. Research into droughts and low flows is certainly needed, to address the complex inter-relationships among their areal extent, intensity, frequency, and duration. Medium to long-term drought forecasting will become more important, as New Zealand's water resources are placed under increasing pressure for development.

Acknowledgments

Drs Paul Mosley, George Griffiths, Alistair McKerchar and Tim Davies are thanked for constructive reviews. Ms Kathy Walter is thanked for figure preparation.

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7
Streamflow and the Use of Models

R P Ibbitt and A I McKerchar

Introduction

Water falling as rain or snow onto the land surface moves along defined natural stream channels to the ocean. The rate of a stream's flow or discharge of a moving stream of water is measured in litres per second (L/s) or cubic metres per second (m^3/s). Discharge measured in New Zealand ranges from less than 1 L/s on small streams in dry conditions to 10,000 m^3/s for large rivers in flood.

Streamflow can be divided into components according to the process by which water reaches the channel. Part of a stream's flow is provided by the slow seepage of groundwater into the channel - this is termed the *baseflow*. Some rain seeps into the soil and rapidly moves into the channel through large pores and seepage zones, this is interflow. If the ground surface is not very permeable, temporarily saturated, or even permanently saturated, as is common near channel systems, then rainfall will become flow over the surface into the stream channel - this is saturated overland flow. Interflow and saturated overland flow together are termed quickflow - the rapid runoff of "new" water.

A streamflow model describes the passage of water through a catchment, normally in mathematical terms that can be programmed into a computer. The model links equations that represent each of the important hydrological process in the catchment.

These models are the subject of this chapter.

The Role of Models

The kinds of problems that concern hydrologists fall into two broad categories:

- □ can we describe what happens to rainfall after it reaches the ground, and before it reaches the stream outlet?
- □ can we predict the future response of a catchment to change?

Streamflow models provide tools for responding to these problems.

Models are not scientific theories, though the two are closely linked. A model provides a means for exploring a theory and for testing hypotheses. Thus, streamflow models attempt to describe, explain, and perhaps predict, water movement through a river catchment. To "build" a model requires a comprehensive understanding of the many processes by which water moves, and development of ways to integrate them so that water movement through the entire catchment can be described. They must describe the processes that are operating at a useful level of detail. For example, for studies of drought in a large catchment, total rainfall and runoff at weekly intervals may provide adequate detail, whereas a study of floods may require hourly measurements of rainfall from several gauges and of discharge.

The hydrological cycle is a simple framework or model which illustrates routes by which precipitation which falls from the clouds ultimately returns to the atmosphere to form new clouds. Once precipitation has become water on the ground surface, individual molecules may follow many paths. A molecule may begin by infiltrating vertically into the soil, only to encounter a stone, a root or worm hole that deflects its motion toward the ground surface; once on the surface it may run down slope, or get caught in a puddle and then soak into the ground. No two molecules ever follow exactly the same path. Since it is impractical to follow the path of every molecule, a more pragmatic approach is required if practical problems are to be solved. A number of different approaches are used in devising streamflow models.

An examination of some typical rainfall and runoff measurements (Figure 7.1) conveys some of the issues which hydrologists may address using streamflow models. In December 1982 a storm was recorded in the basin of the Otaki River which drains the Tararua Ranges north of Wellington. The rainfall was measured with a raingauge near the centre of the 306 km^2 catchment, and the runoff is expressed as discharge measured at the catchment outlet divided by catchment area. These

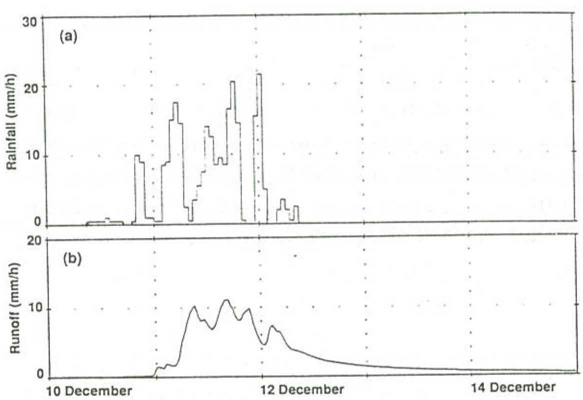

Figure 7.1 **a)** Hourly rainfall depths recorded at a raingauge in the Otaki River catchment for 10-14 December 1982. Total rainfall for the storm is 268 mm. **b)** Runoff recorded for the Otaki River (306 km^2) for 10-14 December 1982. (Runoff is the spatial average for the catchment: it is calculated as discharge (L^3/T) divided by catchment area (L^2)).

quantities are, respectively, inputs and outputs to the catchment system. Study of these data raise all sorts of questions about what happens in a catchment when rain falls:

- ☐ Is rainfall distributed uniformly over the catchment?
- ☐ How representative is the single raingauge of rainfall over the catchment?
- ☐ Why do the runoff rate and rainfall rates differ?
- ☐ If total rainfall was increased by say 20%, how would the runoff change?
- ☐ How would the runoff differ if the storm had occurred at another time of year?
- ☐ If the same amount of rain fell over a shorter period, how would the runoff change?
- ☐ Can patterns be identified that characterise the catchment's runoff response to rainfall?
- ☐ Do these response patterns apply to similar catchments? - That is, would runoff be the same if this storm fell over another catchment with similar characteristics?
- ☐ What are the characteristics which make catchments similar in their response? Can they be measured?
- ☐ At what rate does runoff decline when rain ceases? Can this information be used to predict low flow magnitudes in the absence of rain?
- ☐ What can be said about the soil moisture status before, during and after the storm?
- ☐ How would clearance of native forest from the catchment affect the runoff?
- ☐ How can the runoff rate be deduced from the rainfall rate, and vice-versa?
- ☐ Can the rainfall be used to provide forewarning of two or three hours for flood peaks?

Streamflow models can help to answer these questions. In practice, such models have three principal uses (Clarke, 1973):

Firstly, they are used for **forecasting:** that is for estimating magnitudes of future river flows. **Real-time** records of rainfall and streamflow may be available up to the present time t, and estimates may be needed of future streamflow at times t + 1, t + 2 ... t + k; this is termed operational forecasting. For a hypothetical storm and a hypothetical state of catchment wetness, it may be required to estimate the resultant flood hydrograph; this is flood prediction.

Secondly, where a short discharge record but a long rainfall record are available for a catchment, a model can be used to estimate the flows which would have corresponded to the rainfall, thus extending the discharge record.

Thirdly, models can be used to predict the consequences of possible changes, such as climate change, drainage and river improvement schemes, afforestation, or urban development of previously rural catchments.

The Unit Hydrograph

The unit hydrograph is an early example of a rainfall-runoff model. It is based on considering water drops rather than water molecules. As drops continually break up and reform, so, there would seem to be a poor chance of following the motion of individual drops as though they were marbles. Nevertheless, treating rainfall as a collection of marbles falling on a river catchment has led to advances in the **quantitative** estimation of how long it takes rainfall to arrive at the catchment outlet as streamflow. The unit hydrograph method can be explained in these terms.

The unit hydrograph is the discharge response of a catchment to a given (usually 10 mm) rainfall excess. Rainfall excess means rainfall that runs off quickly and appears as quickflow rather than infiltrating into the soil and appearing as baseflow. The rain is assumed to fall uniformly over the catchment in unit time, typically 1 hour. Figure 7.2a illustrates the unit hydrograph for the Otaki catchment. It is assumed that a catchment responds consistently and linearly to units of rainfall excess: thus a doubling of rainfall excess in unit time will always generate doubled discharge (Figure 7.2b). Also, the assumption of linearity implies that the discharge response from rainfall excess in

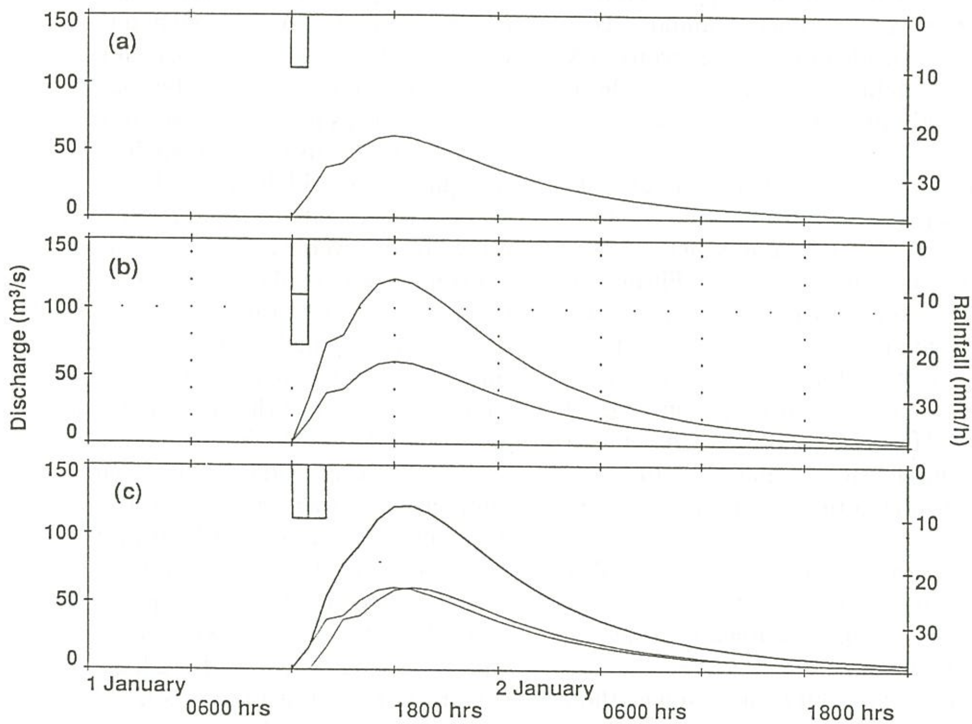

Figure 7.2 The unit hydrograph for the Otaki River. **a)** Unit rainfall excess in unit time (i.e. 10 mm in 1 h), and the corresponding discharge, described as the unit hydrograph. **b)** Linearity means that doubled rainfall excess gives doubled discharge. **c)** Linearity allows hydrographs for separate increments of rainfall to be added to give a single hydrograph.

successive time intervals can be added to provide a total quickflow hydrograph (Figure 7.2c).

This method to predict runoff from rainfall excess has been widely and successfully used for over 50 years. Much of its early success was because it did not require a computer, and could be used even where data on the spatial distribution of rainfall and ground surface properties were inadequate. The unit hydrograph method succeeds because it uses only rainfall and runoff data, and does not consider details of the hydrological processes that operate in a catchment.

Some Principles of Modelling

Because the unit hydrograph was developed for use in engineering design and flood prediction, greater priority was given to developing a method that worked than to developing one based on physical principles. However a physical explanation with attractive mathematical properties is that the unit hydrograph results from the rainfall excess passing through a cascade of linear reservoirs. (A linear reservoir is defined as a reservoir whose discharge is directly proportional to its contents (Nash, 1957).)

This explanation is termed a "conceptual" model, because a particular concept (a cascade of linear reservoirs) is invoked to represent the movement of water through the catchment. Fitting the model requires determination of just two parameters, the number of reservoirs, and the proportionality constant relating discharge to volume in storage. The unit hydrograph method was not developed from such a concept, but it is an early catchment model since it quantifies the rainfall to runoff conversion process in idealised and simplified terms.

Modelling systems can be classified as **distributed** if variations across the catchment are accounted for in the model, or **lumped** if variations are assumed to be negligible and are represented by an average value. The unit hydrograph method is thus a lumped model because a mean rainfall over the catchment defines the input. In contrast, other modelling systems are distributed because variation in rainfall depth across the catchment is accommodated in each system. So, although early catchment models were heavily weighted towards flood prediction, they differed from the unit hydrograph in that they could include information on individual processes. Some processes were described already by theories with attendant mathematical equations, albeit under somewhat simplified conditions. Examples include Philip's (1957) infiltration formula, Penman's (1948) method for calculation of potential evaporation, and Lighthill and Whitham's (1955) kinematic wave routing of flow down steep channels. Empirical equations validated by observation, such as Horton's (1939) infiltration equation, provided alternatives. For groundwater flow to stream channels, the linear reservoir equation provided a satisfactory explanation for the often observed exponential shape of hydrograph recession curves.

The early modellers had to use imagination and intuitive reasoning to deal with phenomena such as the interception of moisture by vegetation or transport of moisture through unsaturated soil. For example, it was known that light rain seldom wets the ground under trees, but that during heavy rain water drips from the leaves to the ground. A simple concept to describe such a phenomenon was quickly and widely adopted. It treated vegetation as a bucket into which rain fell and from which it then overflowed when the bucket was full. This conveniently described the observed delay between the onset of rain and the start of runoff, and it satisfied the idea that a tree canopy has a finite storage capacity. But it left a problem: what was the size of the storage and how could it be measured? Answering such questions provides a challenge even today. Some parameters, like interception storage, can be determined by measurement, but others must be found by trial-and-error. The art of using a model lies in specifying satisfactory estimates for the parameter values.

Mathematical models which consist of a set of equations can be expressed as a logic flow diagram (Figure 7.3). In drawing up such a diagram, errors of logic in the model may be revealed. Any credible model must be able to account for all the water that "entered" the model or that was already present in it, in terms of outflows, evaporation and

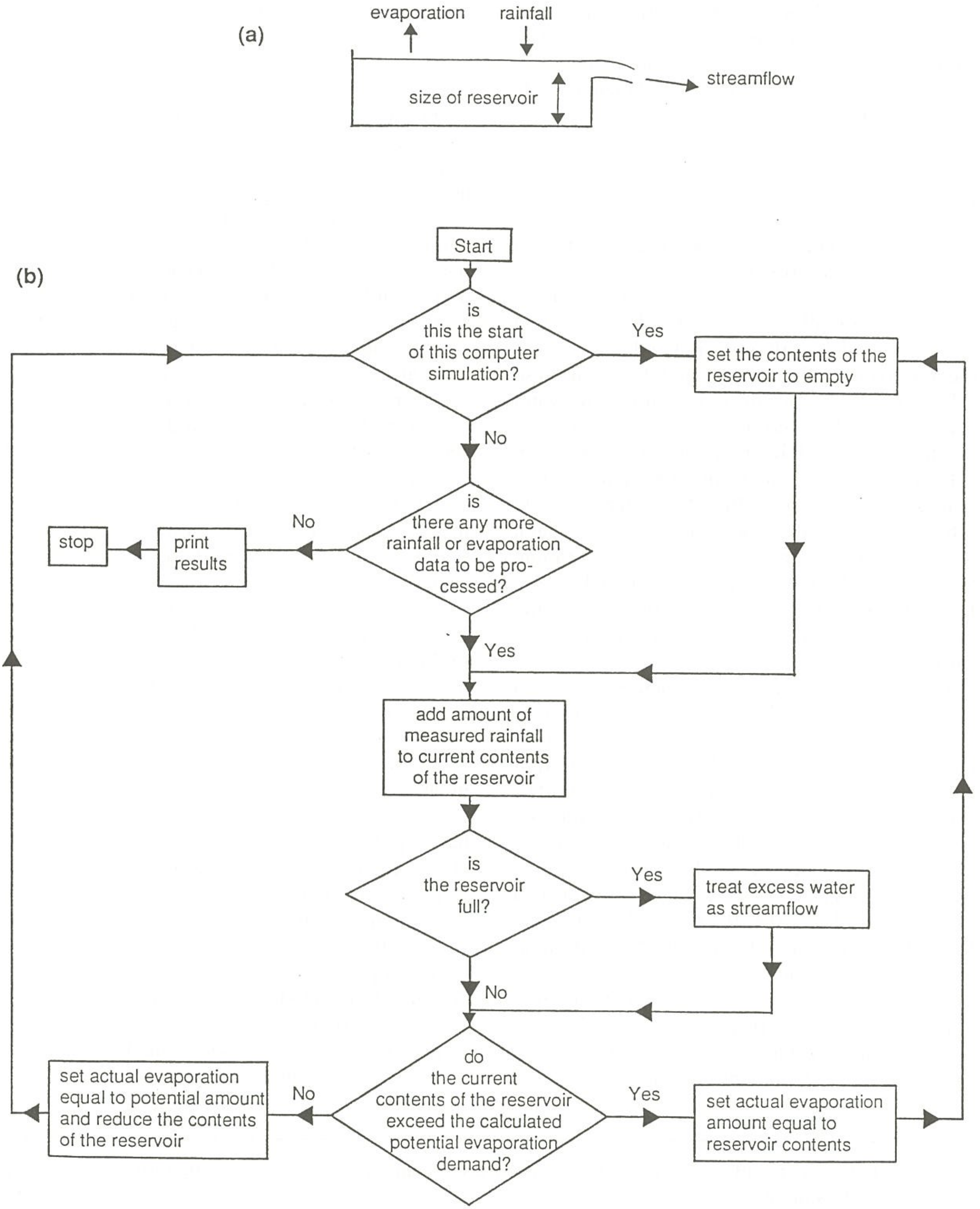

Figure 7.3 Structure of a simple catchment (a) as a logic flow diagram (b).

the final contents of storages. This simple test, known as calculating the water balance of the model, is the expression of conservation of mass in a catchment model. Sadly some catchment models do not pass it.

Data Requirements

To run even a simple model like that illustrated in Figure 7.3(a) requires data. Data come in three forms. Firstly, there are input data, such as records of rainfall over time, which are used to "fuel" the model. Although rainfall data are by far the most common input data, models may also use solar radiation, snowfall, temperature and wind speed to calculate snowmelt, or to calculate how much of the rainfall intercepted by the vegetation gets shaken out. Secondly, there are limiting data that may be measured directly or derived from other measurements. Potential evaporation data, which set a maximum value on actual evaporation, are an example of the latter. The model cannot lose water by evaporation at a rate that exceeds the physical ability of the atmosphere to absorb water vapour[1].

Other forms of limiting data include maximum reservoir contents and maximum infiltration rates.

The third set of data are check data, used to check that the model produces sensible output values. Examples include measured streamflow values and soil moisture variations, or the number of times water ponds on the ground surface. While streamflow data provide a check on the overall performance of a model, other forms of check data increase confidence that the streamflow output is not fortuitously correct but is consistent with what is happening elsewhere in the model.

All too often it is the extent of input data that determines the sophistication of the model used for a particular problem, when in fact it should be the check data that governs this choice.

Guidelines for Modelling

The most taxing problem in modelling is to make a model specific to a particular river catchment. Values have to be assigned to properties such as the capacity of the vegetation to store water, the slope of the hillsides, the slope of the stream channel, or the infiltration capacity of the soil. These values, the model parameters, are assumed to be constant within a river catchment or a pre-defined part of it, but to vary between catchments. Ideally they should be directly measured and should have a physical meaning. In practice, direct measurement is seldom possible and the physical meaning of a parameter is often ill-defined.

Where parameter values cannot be determined directly from field data, trial-and-error fitting procedures can be used. These procedures take initial estimates of the parameters and use them with selected input and limiting data to calculate a model output. The model output data are compared with measured output data, to determine the accuracy of the model. If there is a mismatch, some or all of the parameter values are altered and the process is repeated: a different mismatch will probably result. The process is continued until the model provides an acceptable match with the measured data. The whole process can be done interactively by the modeller, or an automatic procedure may be used which seeks an optimum combination of parameters.

Trial-and-error estimation of parameter values has its pitfalls but many can be avoided by:

- □ using long and varied data sequences to refine parameter estimates;
- □ using more than one type of measurement e.g., river flow and soil moisture data, to assess model performance (Kuczera 1983b);
- □ using simple models or, for more sophisticated ones, restricting parameter refinement to the two or three parameters most likely to influence model performance. The parameters should be

[1] Note that the concept of potential evapotranspiration, while convenient for basin models, has severe limitations (McNaughton, 1976).

chosen which control the volume of runoff generated from the rainfall and the time that runoff takes to reach the catchment outlet. Parameters should be chosen from different processes in the model if possible.

- restraining the impulse to "manipulate more parameter values rather than comprehend fewer" (Kane et al. 1973).

Fitting Models

Although, model parameter measurements are often little more than useful indicators of the catchment-wide value, initial parameter values can be adjusted in order to improve model performance.

Two sets of flows generated by a model are shown, together with the measured flow in Figure 7.4. Which of the two sets of model-generated flows is the better one? A case can be made for either one depending upon the purpose of the simulation. Fit 1 simulates the magnitude of the largest peak much better than Fit 2, and would be the preferred fit if the model is to be used for flood design or flood warning. On the other hand Fit 2 better simulates the hydrograph recessions and also is more responsive to the smaller runoff peaks. This suggests that, with the parameters from Fit 2, the model is behaving more realistically than in Fit 1 and may well give better predictions of unmeasured flows as well as throwing more light upon the processes controlling runoff generation. This presents a dilemma for a hydrologist responsible for recommending which configuration of model is to be used to forecast flood magnitudes for flood warning.

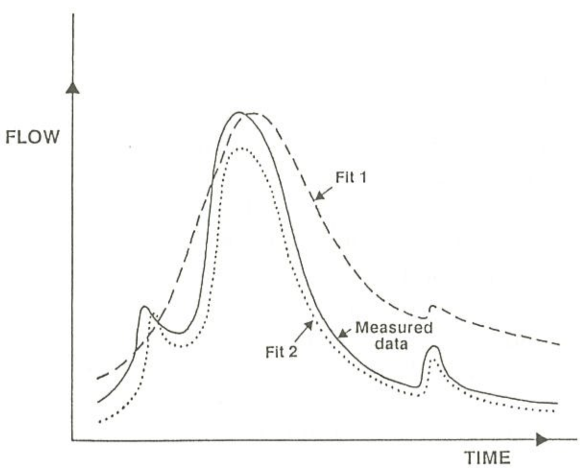

Figure 7.4 Measured flow hydrograph and two simulated hydrographs. The hydrograph labelled fit 1 fits the measured peak well, whereas that labelled fit 2 fits the overall shape of the hydrograph well, but the peak less well.

Clearly there is scope for many different fits, some better than others. For flood forecasting, where lives may be threatened, the best possible forecast is required, but subjective judgement is unsatisfactory, particularly in an emergency. To resolve this problem, automatic means may be used to generate and objectively compare alternatives. Dawdy and O'Donnell (1965), in their pioneering work on the automatic fitting of catchment models, used the sum of squared differences between measured and simulated flows for each measured data point as an objective criterion for selecting the optimum set of parameter values. This fitting criterion is analogous to the sum of squares used in linear regression analysis. It attaches some weight to every recorded flow value; it can be calculated directly from simulated and measured data; for a perfect fit it attains a value of zero. By using every value in the record the fitting criterion is also robust to measurement errors.

The sum of squares criterion has two limitations: first, it gives undue emphasis to high flows (Ibbitt, 1991); second, it produces a time series of differences between measured and computed flows (i.e errors) that commonly are serially correlated, because successive values are similar to the ones that preceded them. These differences therefore contain information that is not incorporated into the model (Clarke, 1973).

Further progress has been achieved by using a maximum likelihood approach to fit an autoregressive moving-average model to the error series (Sorooshian and Dracup, 1980; Kuczera, 1983a), and by incorporating this into automatic fitting procedures (Ibbitt and Hutchinson, 1985). Recent developments include ways to fit more than one type of data series e.g. soil moisture as well as discharge (Douglas et al., 1976 Kuczera, 1983b); and use of a fitting criterion for data at irregular time intervals (Duan et al., 1988). This allows the use of short time steps in calculations when discharge changes quickly, and longer time intervals during low flow periods when flows change slowly.

How are fitting criteria used in practice? The modeller needs to take account of:

- the purpose of the investigation and the level of accuracy required;
- the time by which a result is required;
- the available data for fitting the model;
- the available computing resources;
- previous knowledge of a particular model's behaviour.

A typical modelling strategy might be to:

a) make initial estimates of parameter values from maps, soil surveys, analysis of flow data and/or past experience;
b) do perhaps 10 simulations in which the main parameters are given a range of different "ball-park" values;
c) run an automatic fitting procedure starting from each of the 2 or 3 best sets of values;
d) check to see if the final values for each run are converging to the same set of parameter values or are diverging;
e) compare the model's results with what is known to be realistic and reasonable.

If step (e) suggests that the operation of the model seems unrealistic, return to step (b) and try again. If results are still unrealistic recheck the input, limit and check data. If the data are satisfactory, use a new or longer set of input, limit and check data and/or some new data, e.g., on groundwater levels. Caution is required when new data are introduced (Kuczera, 1983b), to ensure that they illuminate the difficulties being experienced.

Validating Models

If a model is to simulate situations for which no data are available, it should demonstrate that it will produce realistic results. This process, model validation, is perhaps the single most important aspect of a model study. In any study involving the use of a model, some data should be set aside for validation, to independently check that the model generates realistic flows under circumstances different from those used in the fitting process.

Leong et al., (1992) and Woods (pers. comm) used several approaches to validate the models. To assess the potential impacts of climate change on flow in the Hutt River, Leong et al. compared hydrographs generated by the model at internal points in the catchment with measured sub-catchment flows. With hourly data to fit parameters and flows simulated at 4-hourly intervals, Woods demonstrated that his parameter values were not greatly affected by the time interval of the simulation. These examples illustrate the need to be inventive when validating a fitted model, although model type is an important determinant of what can be achieved.

Some Particular Models

Streamflow modelling systems that have been used in New Zealand include RORB, SHE, HYCEMOS and TOPOG. All are classified as distributed modelling systems.

RORB

RORB was developed in Australia (Mein et al., 1974). The acronym stands for Rainfall Routing On a Burroughs computer, although it is now available for other computers. RORB was developed for routing floods through river systems, and simulates streamflow during floods.

Modelling of a river system begins by dividing it into, typically, 10 to 15 sub-catchments. The flow from each sub-catchment is then calculated by routing the effective rainfall - the rainfall minus losses into the ground and into the atmosphere through a non-linear storage function defined by:

$$S_t = K_c K_r q_t m \qquad (7.1)$$

where:

S_t is the storage in the sub-catchment at the start of interval t;

K_r is a relative storage value determined as the proportion of travel time through a sub-catchment reach to the average of travel times through all reaches;

q_t is the discharge during the time interval t;

K_c and m are parameters of the model.

Addition of sub-catchment flows to provide the total catchment response is determined by the layout of the catchment. K_r is estimated from maps and K_c and m are provided by the user following guidelines in the user manual. For many catchments m has a nearly constant value of 0.8.

RORB models can use spatially variable input data and can incorporate the effects of major reservoirs or lakes in a catchment. However, sub-catchments are assumed to be homogeneous. Thus the RORB model for each sub-catchment is effectively a lumped model. RORB models simulate flows at internal points within the catchment but their reliability is not assured. Quality of fit is subjectively assessed and automatic fitting is not included. RORB has been applied to the Hutt River (Figure 7.5) to predict discharge for various design storms. After the parameters had been estimated, the model produced satisfactory validation results (Figure 7.6), and performance for other upstream sub-catchments to which it was not explicitly fitted was also acceptable (Figure 7.7). The model is useful for estimating flood hydrographs corresponding to design rainfall

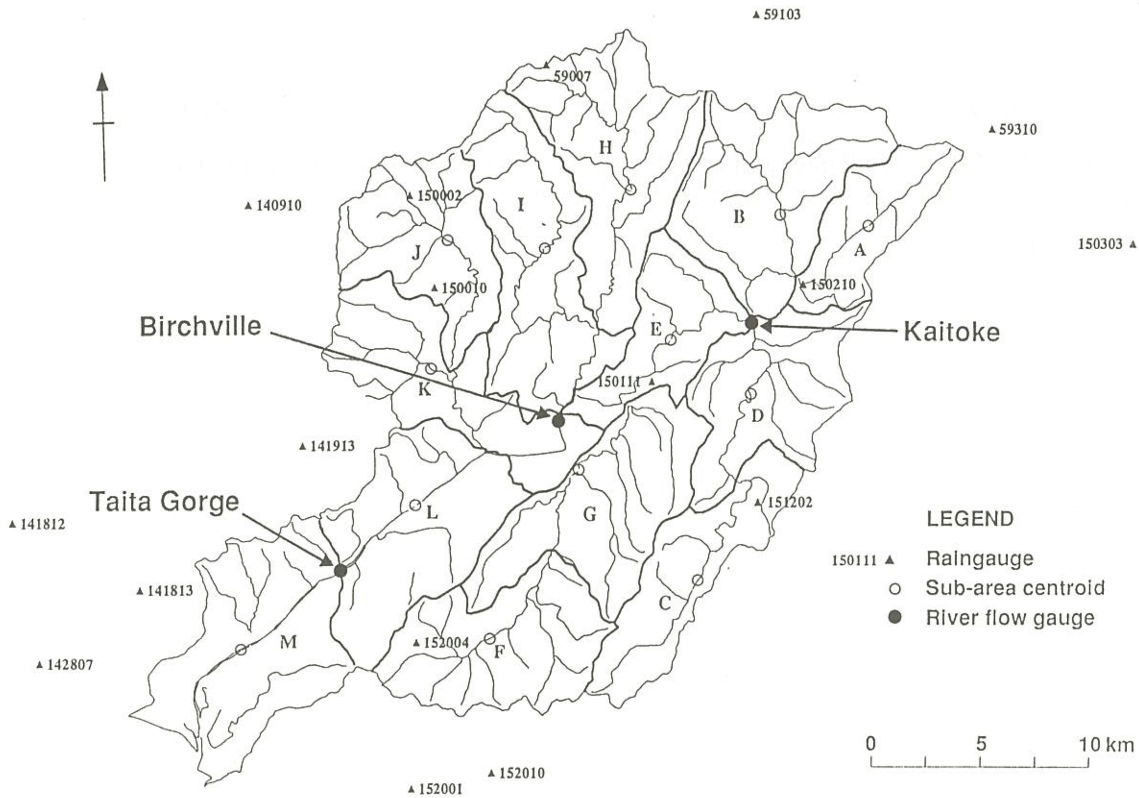

Figure 7.5 Definition of sub-areas (labelled A to M) used in fitting the RORB model to the Hutt catchment. The RORB model was fitted to hydrographs recorded for the Hutt River at Birchville, (e.g. Figure 6). Modelled hydrographs at upstream and downstream locations (Kaitoke and Taita Gorge respectively) are compared with the recorded hydrographs in Figure 7.

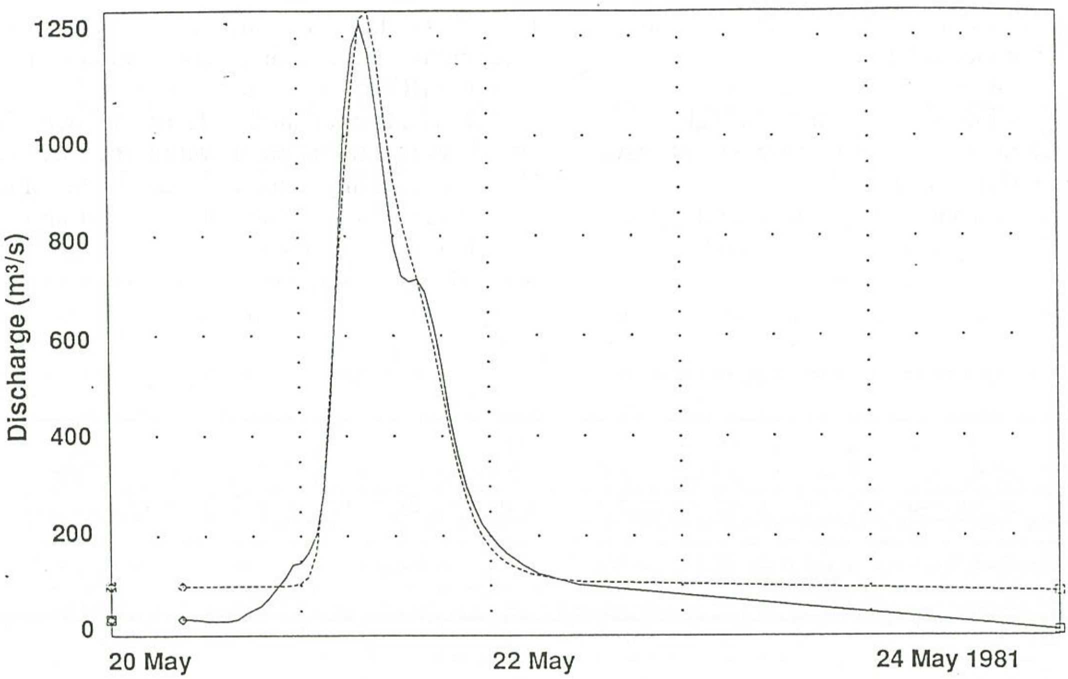

Figure 7.6 Observed flow (full line) and RORB modelled flow (dashed line) for a flood of the Hutt River at Birchville. Corresponding flows at upstream and downstream locations are shown in Figure 7.

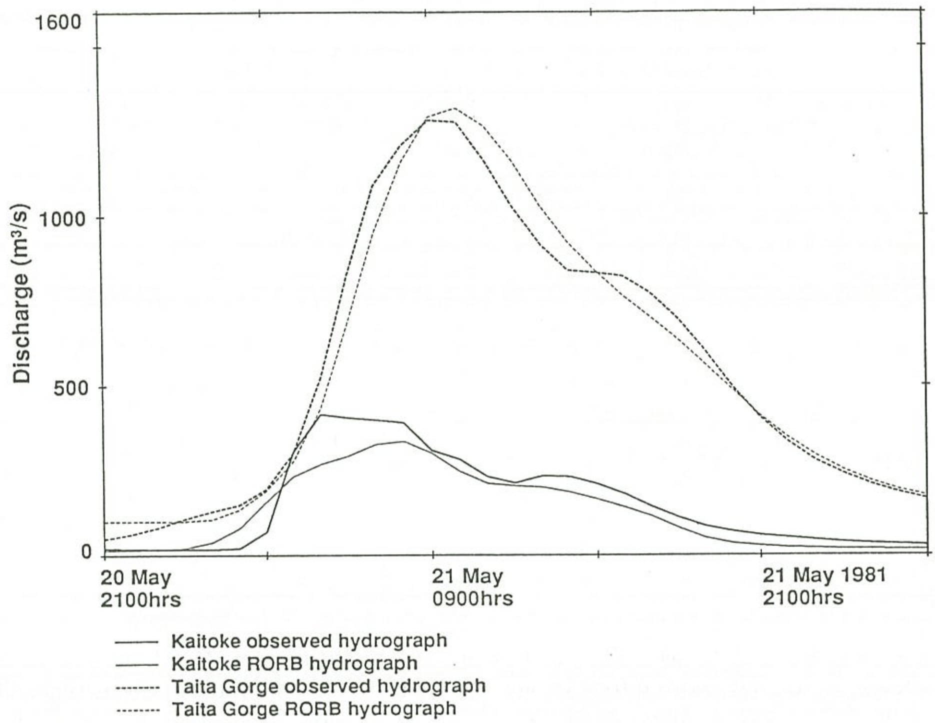

Figure 7.7 Kaitoke and Taita Gorge and observed hydrographs and RORB computed hydrographs for the event fitted in Figure 6.

depths for both urban and rural areas.

In RORB, only quick response runoff from storm rainfall is modelled by a group of storages. Rainfall that infiltrates the ground and appears later as baseflow is treated as a "loss" which is not modelled. More sophisticated models include these slow response components of runoff.

SHE

SHE, the Systeme Hydrologique Europeen, is a catchment modelling system in which the equations of water movement for each of the main hydrological processes within a catchment are solved. Often the equations are too complex to solve exactly, and successive approximations that converge towards solutions are used (Abbott et al. 1986a, b). Calculations of flows between some model components are linked to ensure internal consistencies. For example, seepage from the unsaturated soil to the groundwater is calculated so that the water table rise is consistent with the amount of downward seepage. In SHE, the catchment being modelled is subdivided by overlaying a rectangular grid. The equations for water movement are applied to each grid square. Typical dimensions of the grid squares are 100 m by 100 m, so many elements are required to simulate large catchments. A typical time step for calculations is 3 minutes. Because of the fine subdivision of the catchment a great deal of calculated information is available to compare with measurements made within the catchment. The system's main drawbacks are:

- □ it requires considerable data collection and preparation work before simulations can be run;
- □ the gridding scheme is not easily changed, although later versions of the software are more user-friendly (Nathan, pers. comm.);
- □ the calculations require large computers and large amounts of computer time;
- □ automatic fitting is impractical as well as being inconsistent with the model philosophy;
- □ parameter values must be estimated for many grid points since measured data are not available and this creates uncertainty in the results.

Figure 7.8 shows results obtained for a SHE model of the 38.9 ha Pukewaenga catchment in

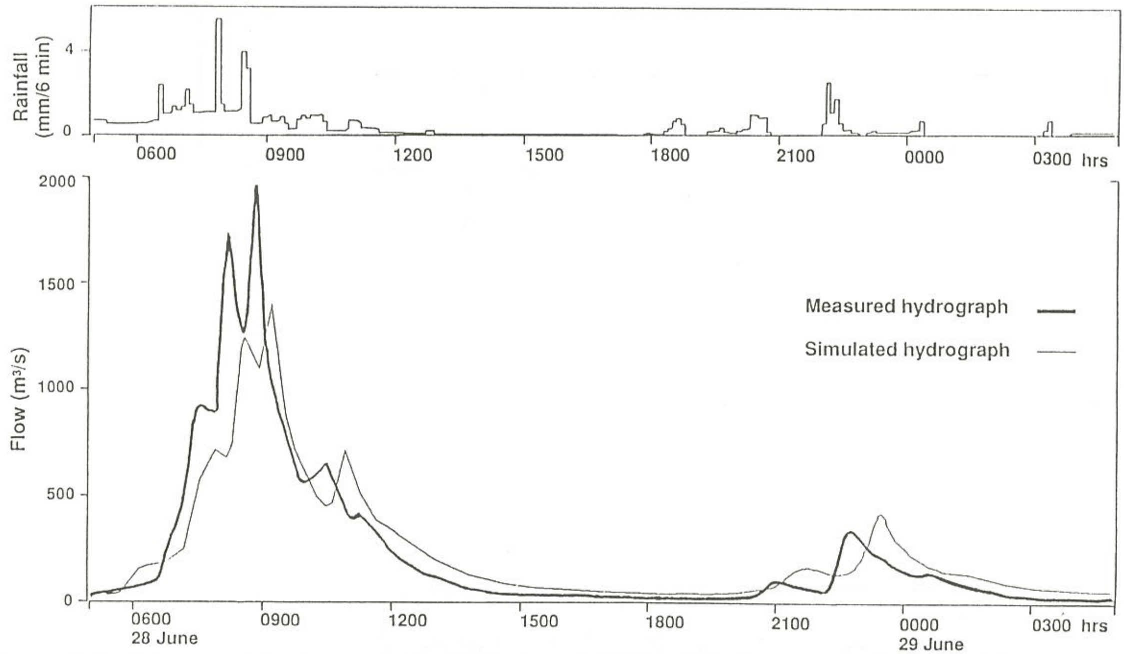

Figure 7.8 Measured hydrograph (bold line) and SHE validation output flows for the Pukewaenga catchment in Northland, New Zealand.

Northland, New Zealand, while Figure 7. 9 shows the arrangement of grid squares used. The simulations for this catchment did not use the sub-surface components of SHE since the flood runoff is largely surface-derived.

HYCEMOS

HYCEMOS, the Hydrology Centre Modelling System, aims to avoid the constraints of numerical models. It is flexible, able to simulate spatially diverse catchments, and uses realistic analytical solutions. Models using numerical solutions have spatial sub-divisions and time step lengths dictated by computational considerations rather than catchment characteristics. New Zealand's relatively steep terrain and fast-flowing rivers are well approximated by kinematic wave theory which says that in "steep" rivers, a flood hydrograph propagates down a river with a gradually steepening rising limb, and without attenuation. General analytical solutions have been developed to apply to kinematic waves (Woods & Ibbitt, 1988). The

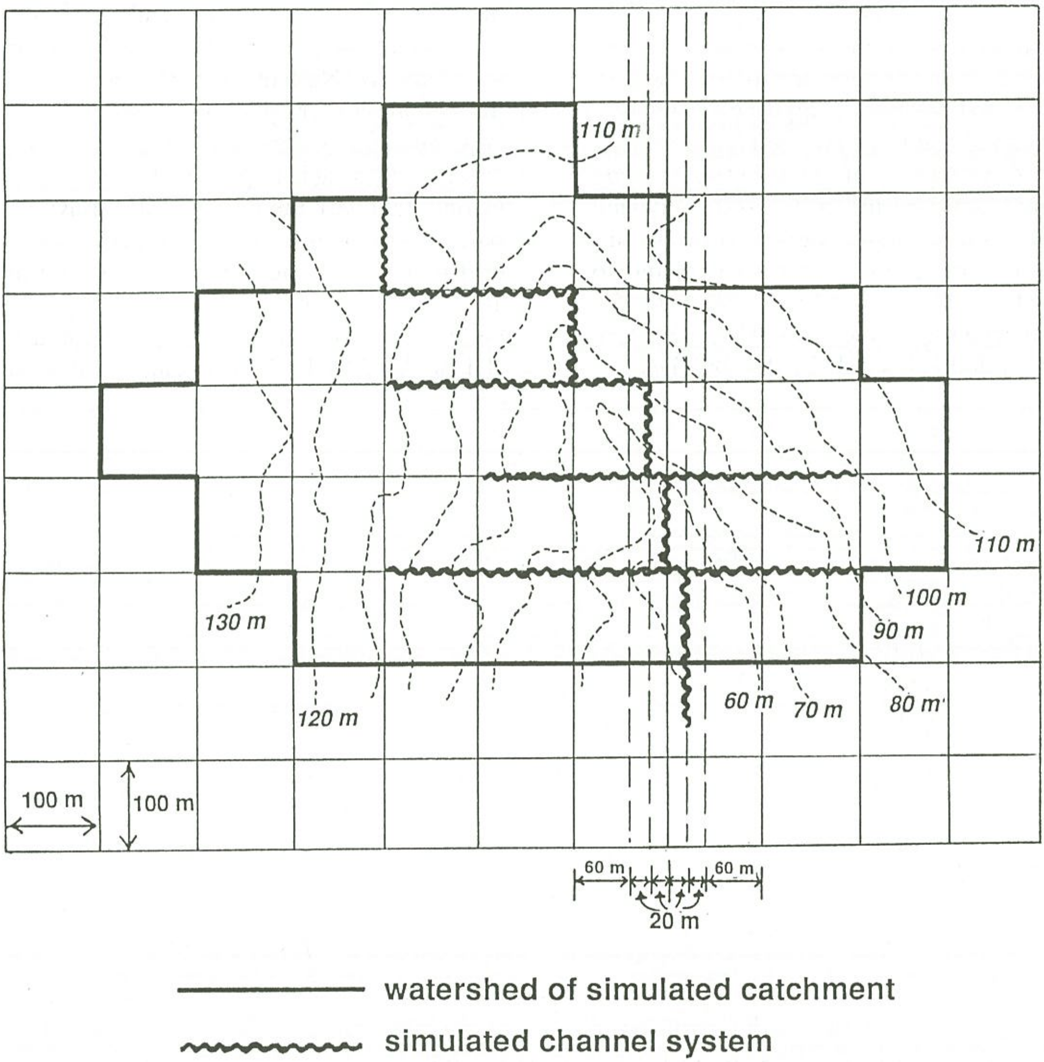

Figure 7.9 Model grid and channel configuration for a simulation of Pukewaenga catchment.

basic unit of the model is a "hillslope" like an open book (Figure 7.10) which is a good approximation of the shape of many the straight V-shaped valleys to be found in New Zealand. Around 10 different hillslopes usually give a satisfactory simulation for a catchment.

HYCEMOS simulates a shallow water table using a scheme proposed by Sloan & Moore (1984), and infiltration and soil moisture movement is based on a simplification of the piston flow model proposed by Clapp (1982).

The stream channel network is simulated by combinations of gutter units with no upstream inflow but lateral inflow, and reach units that accept upstream inflows but no lateral inflows. (Figure 10.) Flow routed through the gutter units uses the same analytical solution as is used for overland flow. Since there is no upstream inflow to either hillslopes or gutters the kinematic wave solution is unconditionally stable. Flow entering reach sections is also routed using a kinematic wave technique. Checks are made for kinematic shocks and the solution given in Henderson (1966) is implemented when these occur.

Hillslope units typically range from 10 m to 1,000 m from river channel to the ridge crest, and can be several kilometres in length along the channel.

HYCEMOS can simulate branched river networks, and includes automatic fitting facilities.

Because HYCEMOS uses mainly analytical solutions to physically based equations, it can simulate the flows at variable time intervals. Thus flood hydrographs, during which the discharge changes quickly, can be simulated using short time intervals. As the flows decrease the simulation time steps can be increased to reduce the amount of computer time needed.

HYCEMOS models can easily run on microcomputers, which makes them useful for flood forecasts.

Drawbacks of the system largely stem from the simplifying assumptions needed to develop analytical solutions. For example a hillslope unit

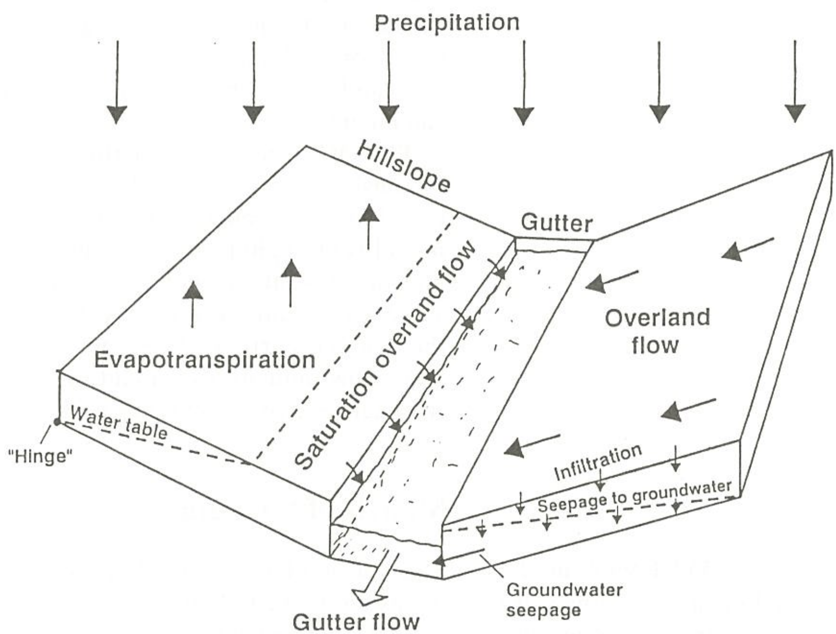

Figure 7.10 A pair of conceptual HYCEMOS hillslopes illustrating the "open-book" idea. Overland flow and channel flow are routed using a kinematic wave technique.

must be assigned a single slope and soil depth, each gutter and reach unit must be straight and rectangular in cross-section, and hillslopes must be rectangular in plan. Rainfall and evaporation are assumed to occur at a uniform rate within each time interval, while infiltration occurs at the rainfall rate or the saturated hydraulic conductivity, whichever is the smaller. The properties of the soil, gutters and reaches are assumed to be constant within each hillslope, gutter or reach. These assumptions are not unduly restrictive and can be overcome by using more units in the model.

HYCEMOS has been used to model the Tahunaatara River (Figure 7.11), a tributary of the Waikato River in the central North Island. The soils of the catchment are very absorbent and the shape of the topography is markedly different from the idealised open-book concept (Figure 7.10). Figure 7.12 compares two validation results for the model with recorded discharge: one validation run uses the same calculation time interval, 1 hr, as that used to fit the model; the other uses the same parameter values but time intervals of 4 hours. These results demonstrate that the model performs acceptably well with varying time intervals.

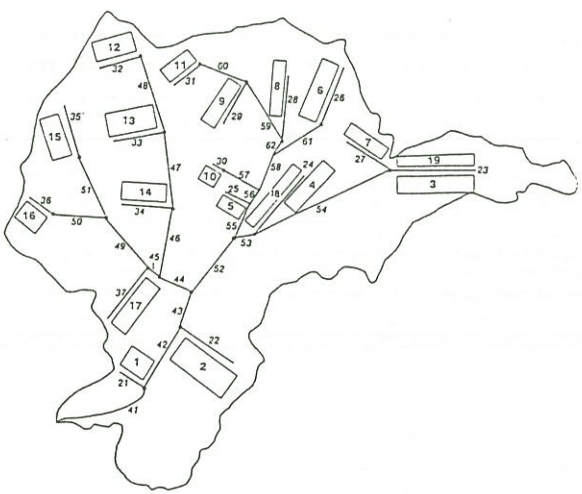

Figure 7.11 Illustration of a HYCEMOS model used for Tahunaatara catchment. Hillslopes are indicated by rectangles (for clarity, these are not drawn to same scale as map), with a bold number to identify the hillslope. Gutters (labelled 21-37) and reaches (labelled 41-62) are indicated by lines with italic numbers.

TOPOG

The fourth model, TOPOG, works on a rather different principle than the three described so far. The latter all include some information on the topography of the catchment, but do not give it a central role. RORB uses topographic information to sub-divide the catchment and calculate the routing coefficients K_r. SHE needs elevation data to determine the mean height of grid elements and channel sections, and HYCEMOS needs approximate slopes for hillslopes and channel sections.

TOPOG, however, uses digital elevation information to define "flow strips" of hillslope which are bounded by flow streamlines, such that rain falling between streamlines always follows the same path, normal to the contours, down the channel (Figure 7.13). Subsurface flow is assumed to be parallel to the catchment surface. Equations based on soil physics describe subsurface flow rates and the soil moisture content throughout the catchment.

TOPOG computes equations of flow in one dimension only, a valuable simplification compared with grid cell methods. Early versions of the model (O'Loughlin, 1986) simulated steady-state moisture conditions within a catchment and took slope aspect into account. A dynamic version of the model is currently being developed which will allow flow simulations, and different, and perhaps stronger tests of the system's performance.

Models of the Future

The four models outlined have been used to varying extent in New Zealand. There is a continuing need for improved models that provide better forecasts and predictions so that timely reliable warnings can be issued, designs improved, and impacts of real and potential changes better understood.

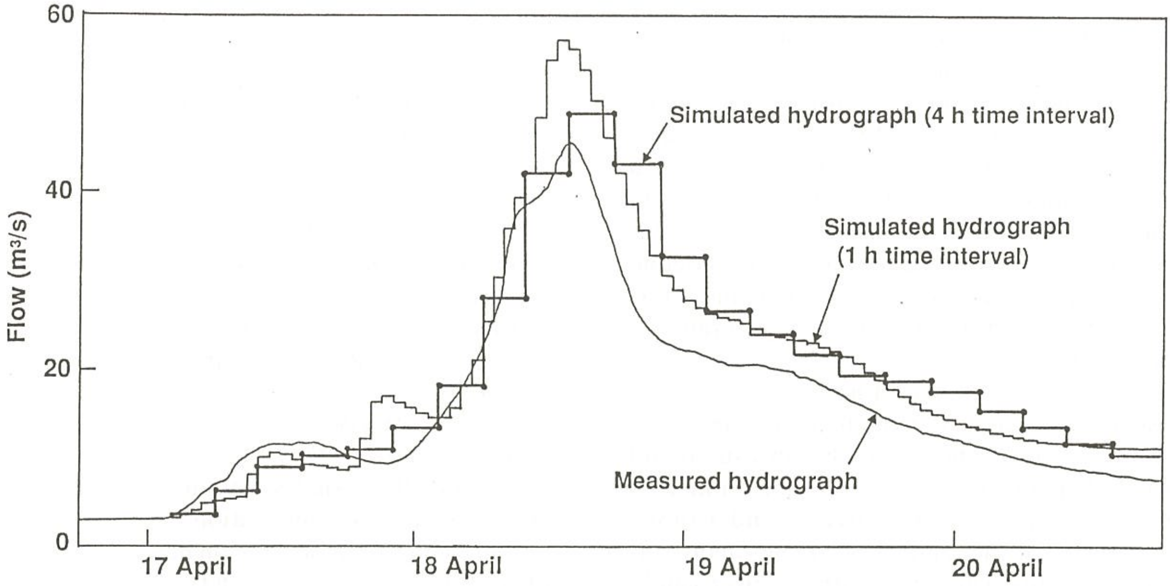

Figure 7.12 Validation simulations for the Tahunaatara catchment at 1-hour and 4-hour time intervals.

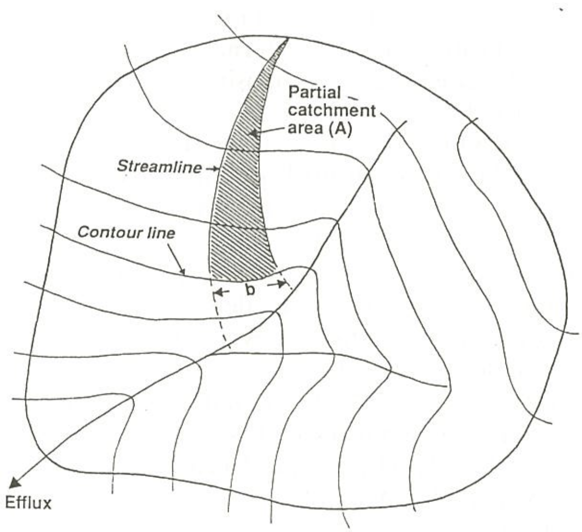

Figure 7.13 Illustration of the development of flow strips to represent a catchment surface in TOPOG. Rain falling anywhere on the cross hatched surface (A) will drain through the contour length b.

There is potential for improvements from use of weather radar, remote sensing and more extensive real-time data collection. There is also potential for development of large-scale models which rest within the grid elements of Global Circulation Models (GCM) to study the effects of climate change at regional rather than continental scales (Shuttleworth 1991). It is unlikely, however, that GCM's will use grid elements with sides smaller than several hundred kilometres (Giorgi & Mearns 1991).

Scale is a pervasive issue in hydrological modelling. Improved modelling requires innovative theoretical and numerical approaches combined with new methods for data collection. Scale determines the temporal and spatial resolutions necessary for useful data collection.

Representative elementary areas have been postulated recently as a size of subcatchment at which the variation in flows between subcatchments reduces to a minimum. Below this threshold, variation between subcatchments increases as the effect of individual features, e.g. a clump of particular vegetation, or a marshy area, predominate. Sampling from subcatchments larger than representative elementary areas would give the same variation in flow response. Representative elementary areas thus may be a basic building block for defining features of a catchment.

To date, representative elementary areas have been demonstrated in simulation experiments using catchment models. The existence of such areas may be confirmed by measuring flow at a range of spatial scales, and by observing the catchment characteristics such as topography, vegetation, soil type and depth, and rainfall which define the elementary areas. A successful outcome of this work will be a significant forward step in rainfall-runoff modelling.

Success will be measured not only by matches between observed and modelled discharge, but also by the matches between other measured and modelled quantities such as evaporation and transpiration rates, soil moisture, and canopy storage. When the simulation of the many different fluxes that can occur within a river catchment closely match corresponding measurements, hydrology will have become a complete earth science.

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8
Evaporation, Soil, and Water

F M Kelliher and D R Scotter

Introduction

In New Zealand most water reaches the land surface as rain. Some rain evaporates after being caught on the surfaces of plants, a process called interception. The rest reaches the ground and usually soaks into the soil, a process called infiltration. Water in the soil is extracted by plants through their roots, and after passing through the stems and branches evaporates through tiny leaf pores known as stomata, in the process called transpiration. Water also evaporates from the moist surface of the soil. The rate at which water evaporates is determined by energy availability, surface wetness, nature of the vegetation, and the availability of soil water.

Evaporation requires a large amount of energy. The net radiant energy from the sun and sky is partitioned into sensible heat that warms the air, latent heat that is consumed in evaporation, and heat that enters the soil by conduction. Latent heat is released when water vapour condenses as precipitation, and is a major heat source in the atmosphere. Thus, for New Zealand, evaporation from the surrounding expanse of sea is a part of the energy balance that contributes to our generally temperate maritime climate.

Evaporation depends on turbulence, with air from aloft continuously mixing with air in the vicinity of the evaporating surface. The irregular surface of plant canopies slows down air flow unevenly, creating turbulence. However, a boundary layer of relatively still air is present just above all evaporating surfaces. Evaporation through this layer is slow compared to the transport of water vapour by convection in turbulent air outside the boundary layer. The boundary layer is thicker in relatively still air, and thins with increasing wind speed and turbulence. The vegetation structure, particularly canopy roughness and height, can thus have a profound effect on the evaporation rate.

Soil, and especially the soil surface, plays a pivotal role in the dynamics of water availability for evaporation. On soils with low permeability, some rain can pond and be lost to streams and rivers as surface runoff. Rain that does infiltrate the soil can have one of several fates. On steep slopes with a very permeable surface layer, such as forest floors or topsoil, water can be lost by rapid, downhill subsurface movement. More commonly, infiltrated water is stored in the soil, until it is either extracted by plant roots or percolates slowly out of the root zone into the subsoil to become ground water. Soil water and ground water intermingle when the water table rises into the root zone, and if the water table rises to the surface, water is lost by surface runoff.

Water retention in soil involves energy relationships. In unsaturated soil, water adheres to soil particles because of surface tension and adsorption forces. Energy is thus needed to extract water from unsaturated soil, although water can flow spontaneously from saturated soil under the influence of gravity. Soil-water movement also depends on a soil's hydraulic conductivity, determined by pore size distribution and water content. Soil-water storage results from the interaction of infiltration, redistribution, and drainage.

This chapter introduces the principles governing the processes outlined above and illustrates them with New Zealand examples. We present models incorporating established theories on evaporation and soil water to introduce some of the essential tools needed for hydrological study and catchment management.

Evaporation from Wet Surfaces

At a basic level, evaporation involves the diffusion of water vapour molecules away from a wet surface. Dalton demonstrated in 1801 that evaporation rate is proportional to the difference between the saturated vapour pressure of the water (determined by water temperature) and the vapour pressure of the air (a function of air temperature and humidity). Evaporation rate is also affected by turbulence, which Dalton called air circulation. Dalton's equation thus illustrates the principles of evaporation, and it is written as (Calder, 1990):

$$E = (e^*_w - e_a)\, f(u)/\rho \qquad (8.1)$$

In this equation, E is evaporation rate (units are m of water s^{-1}), e^*_w is saturated vapour pressure of water (Pascal, Pa), e_a is vapour pressure of air (Pa), f is a function (s m^{-1}) describing the dependence of air circulation (turbulence) on wind speed (u), and ρ is the density of water (998 kg m^{-3}).

Evaporation is a component of the energy balance. In 1948, Penman combined the energy balance with Dalton's equation to calculate evaporation rate without knowing the surface temperature. The Penman equation (Monteith and Unsworth, 1990) can be written:

$$E = E_{eq} + \frac{Dg_b}{\zeta} \qquad (8.2)$$

In this equation, E_{eq} is the evaporation rate obtained in equilibrium with an extensive, homogeneous wet surface via the energy balance (McNaughton, 1976) and it is given by:

$$E_{eq} = \frac{\varepsilon}{\rho\lambda\,(\varepsilon + 1)}\, R_n \qquad (8.3)$$

The two terms ε and λ are properties of water in air and are only weakly temperature dependent; ε is the change of latent heat relative to the change of sensible heat of saturated air (1.27 at 10°C) and λ is the latent heat of vaporisation (2477 kJ kg^{-1} at 10°C, with J being the energy unit Joule). The term R_n is net radiation (units are W m^{-2} where W is the energy flux unit Watt, that is equivalent to J s^{-1}). Net radiation includes solar and reflected radiation during daylight, and longwave radiation exchanged continuously with the atmosphere.

Other terms in equation (8.2) are air saturation deficit D (Pa), an expression of humidity, which is the difference between the saturated vapour pressure and the actual vapour pressure in the air, and g_b (m s^{-1}), which is the boundary layer conductance for water vapour. Boundary layer conductance approaches zero under calm conditions, increases with wind speed, and depends on the roughness of the evaporating surface. The final term in equation (8.2) is $\zeta = \rho(\varepsilon + 1)G_vT_K$ where G_v is the gas constant for water vapour (0.462 m^3 kPa kg^{-1} $°K^{-1}$) and T_K is air temperature in degrees Kelvin, °K (10 °C = 283 °K).

The relative importance of E_{eq} and Dg_b/ζ in equation (8.2) largely depends on the relative magnitudes of net radiation and boundary-layer conductance. When boundary-layer conductance is small, the thick boundary layer largely isolates the wet surface from the effects of air saturation deficit in the air above, and evaporation rate E approaches the equilibrium evaporation rate E_{eq}. Alternatively, the energy for evaporation may be advective, that is, it is brought by the wind. Relatively dry air can come from the warm sea or it can be brought down from aloft. Both these conditions are common in New Zealand.

Evaporation from Wet Plant Canopies

The evaporation rate from a wet plant canopy depends strongly on boundary-layer conductance, which in turn is affected by the vegetation structure, particularly height. Representative values of boundary-layer conductance for pasture, crops and forest, with closed canopies and nominal heights of 0.05, 0.5 and 20 m, respectively, are 0.01, 0.02, and 0.2 m s^{-1}.

Differences in wet canopy evaporation rate for the three vegetation types can be illustrated using typical daytime values during periods of rainfall for the variables in equation (8.2); 50 W m^{-2} for R_n, 0.1 kPa for D, 10°C for air temperature, and the g_b values given above. Evaporation rates are thus 1.3 mm d^{-1} for pasture, 1.6 mm d^{-1} for crops, and 6.8 mm d^{-1} for forest.

The high rate of wet canopy evaporation for forests reflects the high interception loss in forests. Forest canopies can catch significant quantities of rain, and lose it rapidly because of high boundary-layer conductance. It is this high evaporation rate, rather than canopy water-storage capacity, which determines the amount of rain lost by interception (Kelliher et al., 1992a).

Net radiation has little effect on the rate of wet canopy evaporation from forests; this evaporation occurs at similar rates both day and night (Kelliher et al., 1992a). Air saturation deficit and boundary-layer conductance are more important. Even so, rainfall interception loss will be proportionally highest during low-intensity, intermittent storms, when evaporation between rain falls allows the canopy to be re-wetted a number of times.

Interception loss was measured as 26 to 39% of rainfall in a range of native forests (Aldridge and Jackson, 1973; Rowe, 1979), and as 12 to 49% of rainfall in *Pinus radiata* plantations (Kelliher et al., 1992a). Interception loss was higher in the taller and more closely spaced stands, and thinning and pruning stands can significantly reduce interception loss (Whitehead and Kelliher, 1991a).

In contrast, interception loss is generally lower in pastures and crops (McNaughton and Jarvis, 1983). An interesting exception is tussock grassland, which generally is located in exposed windy habitats. Though tussocks are relatively short, the level of windiness gives tussock grassland boundary-layer conductance values similar to those of a forest. Interception loss was 21% of rainfall over a year for narrow-leaved snow tussock at Glendhu, south of Dunedin (Campbell and Murray, 1990).

Evaporation from wet soil

Evaporation from wet soil surfaces can also be estimated using Penman's equation. Boundary-layer conductance for bare soil is similar to that for pasture. On a sunny summer day, evaporation may reach 6 mm d^{-1} from a wet field of fallow soil. Evaporation from wet forest floor can account for up to 20% of total evaporation on fine days in mature native forest (Kelliher et al., 1992b). However, during rain-free periods a dry surface layer soon develops on bare soils and forest floors, lowering the conductance for water vapour diffusion, and substantially reducing evaporation.

Transpiration From Plant Canopies

As well as through interception loss from canopies, plants also contribute to total evaporation by transpiration through leaf stomata. Plants can control stomatal opening to influence transpiration rate. Primarily, this process is regulated by soil-water availability, with stomata closing as soil dries via a plant physiological response.

By including a stomatal conductance, Dalton's equation can be extended to describe transpiration from a single leaf. However, here we are interested in transpiration from whole plant canopies. In 1965, Monteith extended Penman's equation so that it could describe transpiration from a plant canopy by using a single bulk canopy conductance g_c. This represents the product of a representative stomatal conductance (g_s) and a dimensionless leaf area index, a_t, which is canopy leaf area per unit ground area.

The so-named Penman-Monteith equation may be written (McNaughton and Jarvis, 1983) as:

$$E = \Omega E_{eq} + (1 - \Omega) E_{imp} \qquad (8.4)$$

where the dry canopy evaporation rate E imposed by the effects of the air saturation deficit D, is:

$$E_{imp} = \frac{Dg_c}{\rho G_v T_k} \tag{8.5}$$

and the coefficient Ω indicating the degree of coupling between the canopy and D is:

$$\Omega = \frac{1 + \varepsilon}{1 + \varepsilon + \frac{g_b}{g_c}} \tag{8.6}$$

The Penman-Monteith equation describes how the relative importance of radiative and advective energy for transpiration depends on the ratio of the boundary-layer and canopy conductances.

When boundary-layer conductance is much less than canopy conductance, Ω approaches 1, and evaporation rate approximates the equilibriumevaporationrate. This tends to be true for low vegetation, such as pasture and many crops, provided soil water is adequate. It is the basis of the widely used Priestley-Taylor evaporation equation (Priestley and Taylor, 1972)

$$E = \alpha E_{eq} \tag{8.7}$$

The empirical factor α is commonly found to be 1.3, which indicates the input of some advective energy, inevitable under most conditions.

Transpiration from Pasture

Transpiration from well-watered pasture will vary from season to season, reflecting the changing intensity of solar radiation (Figure 8.1). At a site in the Waikato Valley near Hamilton, long-term

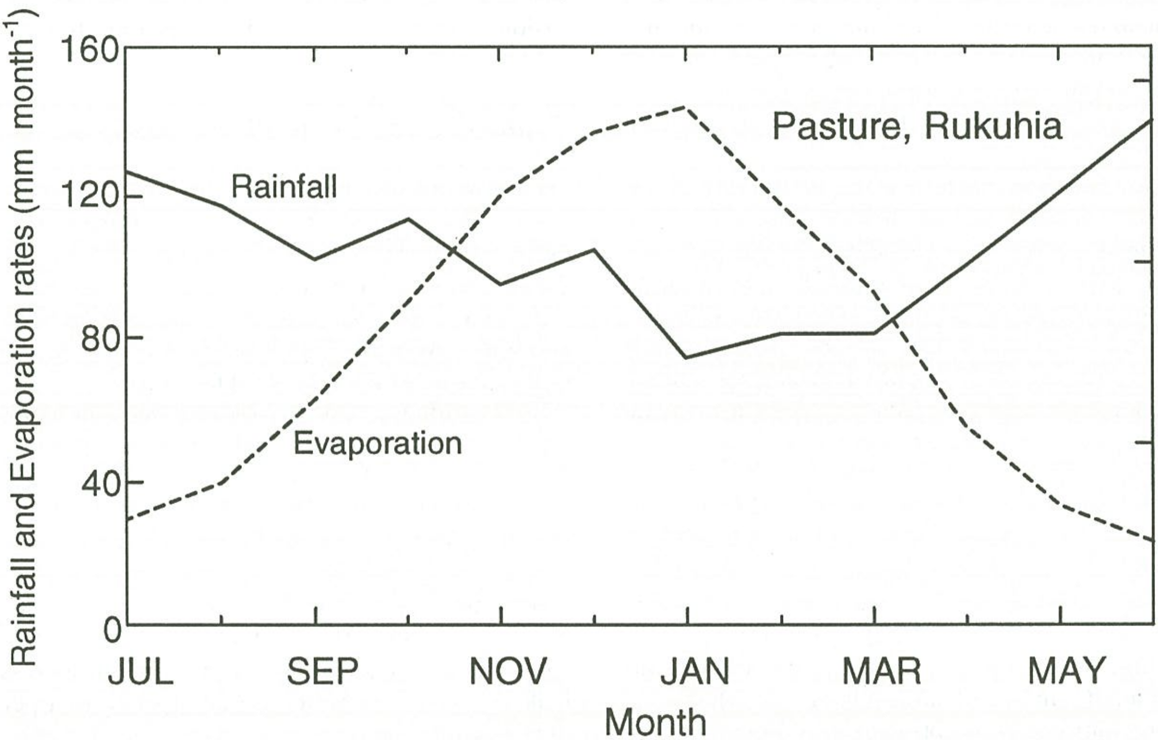

Figure 8.1 Thirty-three year (1946 - 1979) averages of monthly rainfall (———) and calculated evaporation (- - - - -) rates, using the Penman equation as described in the text, for pasture at Rukuhia near Hamilton (after McAneney et al., 1982, with permission). Annual rainfall and evaporation were 1242 and 933 mm, respectively

average rain exceeds total evaporation by about 300 mm per year (McAneney et al., 1982). However, as about 40% of water extraction by the roots of the pasture plants comes from the top 250 mm of soil, fairly regular rain is needed to maintain transpiration and production (McAneney and Judd, 1983). Transpiration rates of about 4 to 5 mm d^{-1} often exceed the rainfall in summer, and soil-water reserves can become depleted. Eventually canopy conductance will decrease, and pasture evaporation will drop to 1 or 2 mm d^{-1} as a water deficit sets in. Growth rates will decrease correspondingly.

In winter, when days are short and solar radiation is minimal, rainfall regularly exceeds total evaporation. Excess water will drain beyond the root zone, often leaching solutes such as nitrates added as fertiliser.

Transpiration from Forest

Forests behave quite differently from pasture and short crops because of their taller stature and adaptation to conditions of periodic water deficit. For such vegetation, boundary-layer conductance usually exceeds canopy conductance, so that Ω approaches zero. Evaporation rate then approximates that imposed by the air saturation deficit, E_{imp}. Here, in the absence of soil-water deficit, stomatal behaviour, leaf area, and interaction with the aerial (advective energy) environment will largely control evaporation. However, air saturation deficit and net radiation are not entirely independent; solar radiation warms the air, thereby increasing the air saturation deficit.

For *Pinus radiata,* in Kaingaroa Forest south of Rotorua, stomatal conductance is greatly affected by air saturation deficit (Figure 8.2(a); Kelliher et al., 1992a). When air saturation deficit is less than 0.5 kPa, stomatal conductance, and hence canopy conductance, is constant and tree canopy transpiration rate is approximately proportional to air saturation deficit. However, for higher air saturation deficits, transpiration decreases because of stomatal closure and reduced canopy conductance.

Tree leaf area, and to a lesser extent tree height, also have a marked effect on transpiration rate, which increases with increasing leaf area (Figure 8.2(b)). Thinning and pruning of the trees can reduce transpiration as well as interception losses, thus increasing water yield from a catchment (Whitehead and Kelliher, 1991a).

For mature forest, tree transpiration is a large component of the catchment water balance (Figure 8.3). However, its contribution may differ for different forest types. Annual transpiration rate was estimated as 700 mm, about 50% of rainfall, for a highly productive *Pinus radiata* forest under

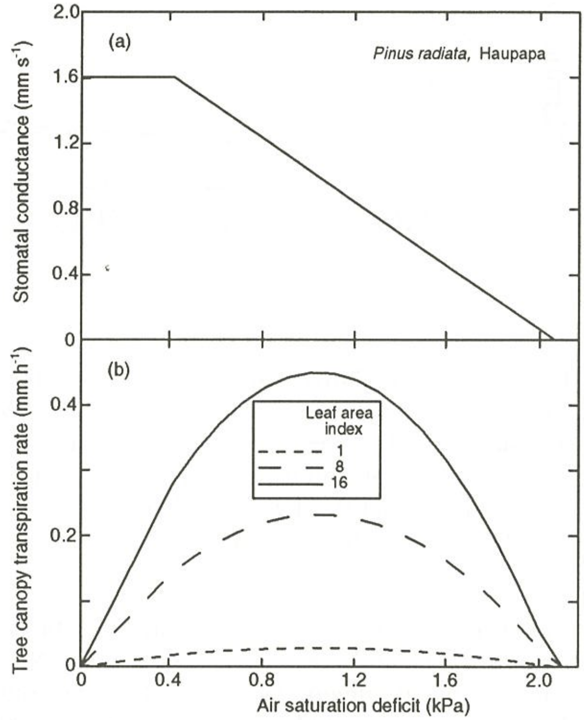

Figure 8.2 The relationship between stomatal conductance and air saturation deficit for Pinus radiata (a) at Kaingaroa Forest near Rotorua (after Kelliher et al., 1992a, with permission). This relationship was used to calculate tree canopy transpiration rate (b) using thePenman-Monteith equation, with $E = E_{imp}$ as described in the text, for three leaf area indices (a_t) approximating a young, open (a_t = 1,-----), an intermediate (a_t = 8, — — —), and a closed canopy forest (a_t = 16, ———).

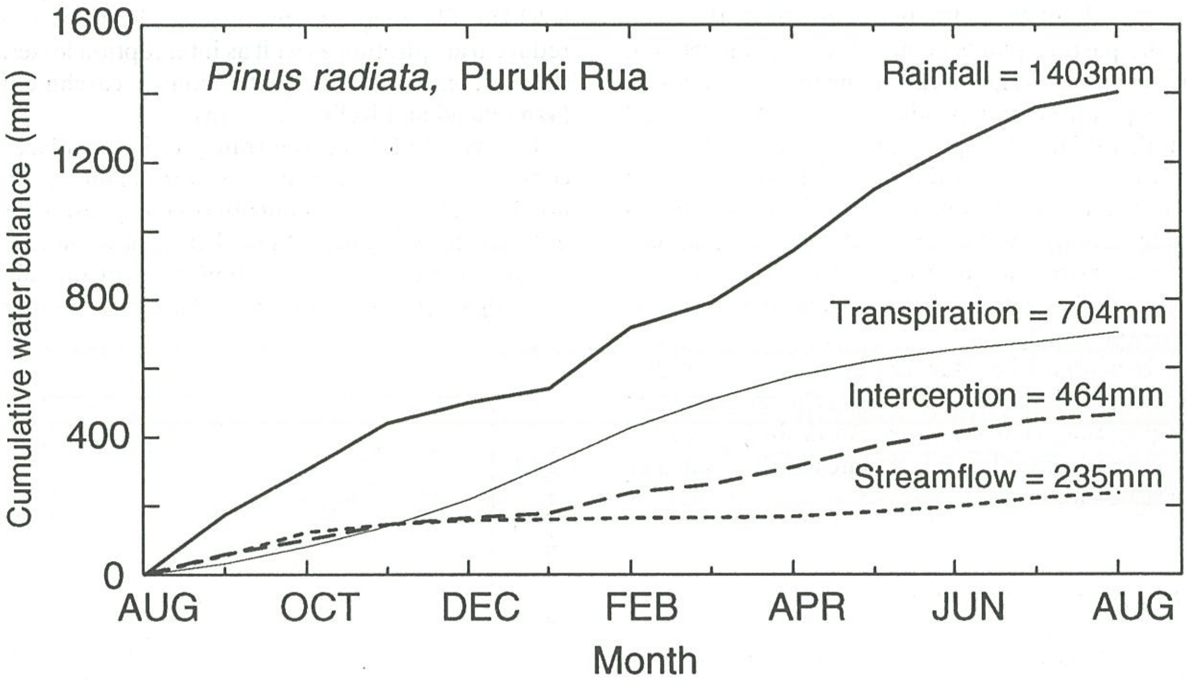

Figure 8.3 Cumulative water balance components for one year in the 9 ha Puruki Rua Pinus radiata catchment, 30 km south of Rotorua (after Whitehead and Kelliher, 1991b, with permission): measured rainfall (———) = 1403 mm, measured streamflow (-----) = 234 mm, tree canopy transpiration calculated by the Penman-Monteith equation (———) = 704 mm, and tree canopy interception calculated as a residual (— — —) = 464 mm. The closed canopy forest was 13 years old and tree leaf area index was 16.

well-watered conditions in the central North Island (Whitehead and Kelliher, 1991b). For a Nelson beech catchment, however, Benecke and Evans (1987) estimated tree transpiration as 420 mm, only 28% of annual rainfall.

Transpiration is also a major component of the water balance for tussock grassland. Campbell and Murray (1990) obtained a value of 400 mm, 38% of annual rainfall, for transpiration from Otago snow tussock grassland.

Soil and Water

Along with energy availability and vegetation type, soil conditions can greatly influence evaporation rate and the water balance. The primary mineral and organic particles of soil can vary in size from large stones and roots down to colloidal clay and humus particles of less than 2 μm in size. Primary particles are usually bound together into larger aggregates. The smaller particles give soil a large and reactive surface area.

The variation in particle size results in a wide range of pore sizes, from large cracks and old root channels to the minute gaps between colloidal particles. The fraction of the soil volume consisting of pores is termed the soil porosity, which is typically between 0.4 and 0.7. As only the larger pores are visible with the naked eye, soils are much more porous than they look.

Soil pores contain a mixture of water with dissolved nutrients, and air, which has less oxygen and more carbon dioxide than in the atmosphere. The volumetric water content, θ, is the fraction of soil volume consisting of water. In a completely saturated soil, free of air, this will equal the porosity.

Soil Water Retention

How is soil able to retain water? Why do soils contain some water that plants cannot extract? To answer these questions we need to go beyond the soil-water content. Free water is present only in effectively saturated soil, where it can spontaneously flow out of the soil under the influence of gravity. However, energy is needed to extract water from unsaturated soil, to oppose both the surface tension and the adsorption forces that cause water to adhere to the soil particles. Surface tension causes air-water surfaces to behave as if they have a skin on them. The result of surface tension, and the strong attraction between water and most solid surfaces, is capillary forces that hold water in the soil pores.

The soil-water pressure created by these capillary forces is less than atmospheric pressure. The pressure potential, P, is a measure of this pressure and of the related energy of soil water relative to free water. Below the water table, where the soil is saturated and free water is present, pressure potential is positive; but in unsaturated soil above the water table it is negative because the water has less energy than free water. We will use head units of metres for pressure potential, which corresponds dimensionally to energy per unit weight of water (1 m $\approx$ 10 kPa).

As a soil dries the larger pores lose their water first, as the smaller the pore the stronger the capillary forces holding the water there. The relationship between pressure potential and volumetric water content thus reflects the pore size distribution. This so-called soil-water characteristic is a fundamental hydrological property of a soil.

The effects of differing soil-water characteristics can be demonstrated by looking at three New Zealand soils (Figure 8.4(a)). Beach sand from Tangimoana has large pores from which water is readily extracted. At a pressure potential of -1 m, this sand has a water content of only 0.02 m^3 of water per m^3 of soil; that is, water takes up only 2% of the soil's volume.

In contrast, Tokomaru silt loam subsoil (from near Palmerston North) consists mainly of wind-blown silt, which has become compacted over time and contains many small pores. Its water content changes only from 0.38 to 0.35 $m^3\ m^{-3}$ as the pressure potential drops from zero to - 5 m because most of the pores in this soil are smaller than 6 μm in diameter.

The alluvial Wairau subsoil from near Blenheim is a fine sandy loam, with some silt and clay. Its soil-water characteristic is intermediate between the beach sand and the silt loam, indicating that it has substantial volumes of both large and small pores.

Soil water is subject not only to the capillary forces included in the pressure potential, but to the downward pull of gravity. Gravitational potential, Z, is simply elevation relative to some reference level when head units are used; only its relative value is important. The hydraulic potential, H, is defined as the sum of pressure potential and gravitational potential:

$$H = P + Z \qquad (8.8)$$

Water in soil moves from higher to lower hydraulic potential, the gradient of hydraulic potential being the driving force. If hydraulic potential is constant throughout the soil, the gradient is zero, equilibrium exists, and there is no water movement.

Soil Water Dynamics

Soil water moves as a viscous laminar fluid. Like honey dripping through a slice of toast, it can flow much more rapidly through large pores than small pores. The two determinants of soil-water movement are the driving force of the hydraulic potential gradient and the soil's hydraulic conductivity, which is determined by the pore size distribution and water content. The equation stating this was

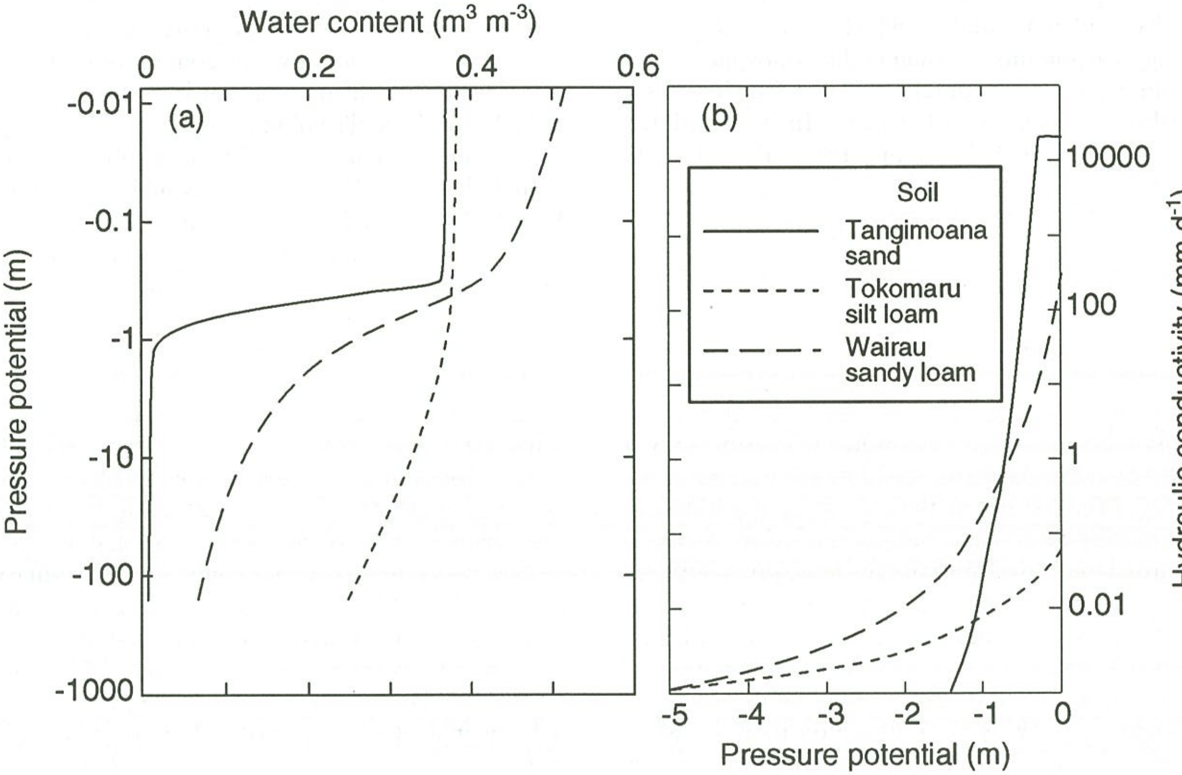

Figure 8.4 Soil-water characteristics (a) and hydraulic conductivity data (b) for Tangimoana beach sand (———), Tokomaru silt loam subsoil (- - - -) and Wairau deep silt loam subsoil (— — —) (J.P.C Watt, personal communication).

first derived for saturated materials by Darcy in 1856. It was extended to unsaturated soils by Buckingham in 1907 (Jury et al, 1991) as:

$$q = -K(P)\ \mathrm{d}H/\mathrm{d}x \qquad (8.9)$$

In this equation, q is the volume flow rate through unit cross-section of soil, K(P) is the hydraulic conductivity at a given pressure potential, and x is distance in the direction of flow. This is a second fundamental hydrological property of soil.

In saturated porous materials, the hydraulic conductivity is independent of pressure potential and is called the saturated hydraulic conductivity, K_{sat}. Unsaturated soil is much less conductive than saturated soil as the larger pores have lost their water, so that hydraulic conductivity varies strongly with water content or pressure potential. Hydraulic conductivity curves for the beach sand and the two subsoils are shown in Figure 8.4(b). The beach sand is very conductive at saturation, pressure potential = 0, with K_{sat} = 21 000 mm d^{-1}. However for the compacted silt loam subsoil of similar porosity, K_{sat} = 0.05 mm d^{-1} because of its smaller pores. The fine sandy loam, with a mixture of large and small pores, has K_{sat} = 311 mm d^{-1}, which is more common for New Zealand soils than the other two values.

The saturated hydraulic conductivity is the maximum sustainable infiltration rate for a thin film of surface water. It occurs when sustained ponding saturates the surface soil. Under these conditions, gravity is the only force involved and the hydraulic potential gradient is -1. The Darcy-Buckingham equation (8.9) then reduces to $q = K_{sat}$. Incipient ponding is relatively rare, as rainfall intensity is usually less than the saturated hydraulic conductivity.

As the pressure potential decreases, becoming more negative, and the larger pores lose their water, the hydraulic conductivity decreases markedly. The conductivity of sand diminishes rapidly once the large pores have drained; few smaller water-filled pores remain, and consequently there are fewer pathways for water movement. When pressure potential is less than -1 m, all three soils have hydraulic conductivity values of less than 1 mm d^{-1}, indicating that water movement through all these soils is very slow at low pressure potentials.

Another key concept needed to describe water movement in soil is the principle of mass conservation. This requires that in any volume of soil over any time period, a difference between the amounts of water going in and coming out will result in a change in soil-water content.

Soil Water Infiltration, Redistribution and Drainage

We will now use the principles outlined to examine soil-water behaviour in three idealised uniform profiles consisting of the sand and two subsoils discussed previously. Extraction of water by plants generally ceases once pressure potential reaches -150 m, when plant roots have difficulty in obtaining the soil's tightly bound water. We use this pressure potential as a starting point in our examples (Figure 8.5).

The three initial water-content profiles differ widely for the three soils. The beach sand contains negligible water, and the Tokomaru silt loam contains nearly 30% by volume of water that is unavailable to plants.

We next introduce 60 mm of steady rainfall over a 6-hour period. The Darcy-Buckingham and

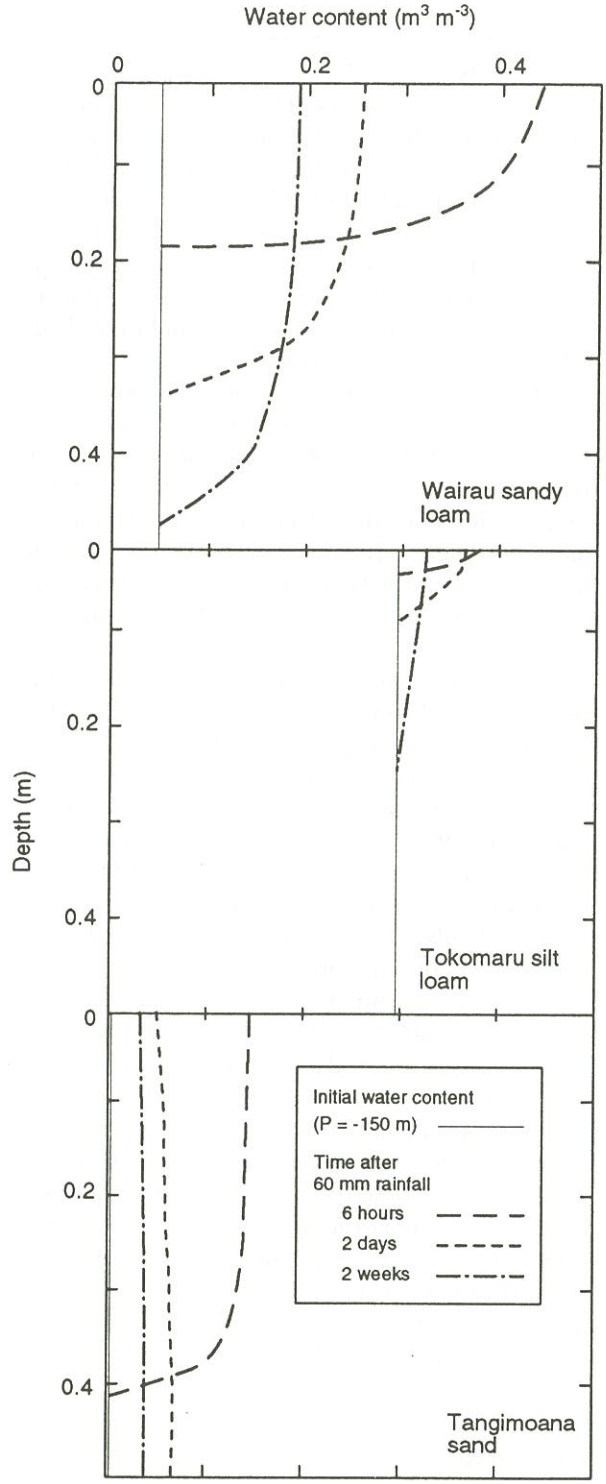

Figure 8.5 Computed water content profiles for the three soils in Figure 8.4 before and after 6 h of rain at 10 mm h^{-1} as described in the text.

mass conservation equations can then be solved numerically (Ross, 1990) to see what happens. The three new water-content profiles calculated immediately after the rain show that all 60 mm of rain infiltrates into the Wairau sandy loam and the Tangimoana beach sand. However, 56 mm of the rain is lost as runoff (with surface detention assumed to be 2 mm) and only 4 mm infiltrates the compacted, less-permeable Tokomaru silt loam (2 mm infiltrates during the 6 h of rainfall and the other 2 mm afterwards)(Figure 8.5).

After the rain, soil water moves downward in response to gravity and capillarity. The calculations assume that there is no evaporation or plant uptake of water. Most of the redistribution takes place in the first 2 days while the soil is wetter and more conductive (Figure 8.5). After 2 days the larger pores lose their water, and subsequent redistribution is much slower (shown after 2 weeks).

All 60 mm of rain remains in the top 0.5 m of the Wairau sandy loam, but only 21 mm remains in the beach sand, which has a much smaller capacity for long-term storage because it lacks smaller pores. As a typical summer evaporation rate is 4 mm d^{-1}, plants growing in the Wairau sandy loam could extract water at that rate for 15 days. But only 1 day's supply has infiltrated into the Tokomaru silt loam, and only 5 days' supply remains stored in the Tangimoana beach sand.

The water retentivity and conductivity characteristics of soils determine their ability to store water for plant use, and allow soils to act as a buffer in the hydrological cycle. The water storage capacity of soils available to plants can be very large; Will and Stone (1967) found a value of over 900 mm for 2.7 m depth of a North Island pumice soil growing Pinus radiata. More typical values range from 45 mm to 173 mm in the top 760 mm of soil (Gradwell, 1974, 1976). The factors involved, and New Zealand research on the subject, are reviewed more fully by Watt (1979).

Conclusions

Evaporation is a dynamic process, determined principally by the state of the physical environment and physical characteristics of vegetation. Evaporation is a major component of the catchment water balance.

Soil plays a pivotal role in linking evaporation into the dynamics of the hydrological cycle. Soil water and hydraulic conductivity characteristics determine soil-water storage and movement. Infiltration, redistribution, and drainage of water are important processes affecting evaporation, surface runoff, ground water, and stream flow.

This synopsis of physical principles and theory facilitates understanding of changes in catchment hydrology that can be the result of changes in climate, land-use, and management practices. Established theory is useful in calculations of evaporation and soil-water storage and movement for purposes of hydrological study and management. These calculations can also contribute reliable quantitative knowledge to planning decisions about water resource management.

Acknowledgements

We are grateful to Drs. Brent Clothier, Paul Mosley, and David Whitehead and Joanna Orwin for providing constructive criticism of our manuscript. We thank Tom Pearson and Tom Farndon for final preparation of the figures. Financial support for the senior author was provided by a grant from the Foundation for Research, Science, and Technology.

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9
Geothermal Hydrology

T M Hunt and H M Bibby

Introduction

Waters which have been warmed by the Earth's natural, internal heat are commonly called geothermal waters. What is meant by warmed and how these waters differ from normal, cold groundwater is difficult to answer precisely. The Resource Management Act (1991) differentiates between geothermal water and groundwater on the basis of temperature: water greater than 30°C is geothermal, that below 30°C is groundwater. This distinction has no scientific foundation; it is based solely on the concept that water above this arbitrary temperature is an energy resource. It is not possible to clearly separate geothermal water from groundwater on the basis of temperature alone.

The major factors which distinguish geothermal waters from groundwater are:

(i) The physical properties of geothermal waters vary significantly, especially with temperature (Table 9.1), and these variations become very large when water (liquid) changes to steam (vapour). For example:
- Hot water is less dense than cold water, and thus is more buoyant. Buoyancy forces are the dominant factor in the movement of geothermal water; for this reason hot, geothermal waters are rarely stagnant.
- Hot water is less viscous than cold water, and thus flows more easily: for the same pressure gradient, water at 100°C flows about four times faster than water at 20°C.

(ii) The chemistry of geothermal waters differs from that of cold groundwater (Table 9.2), and also changes with temperature. Hot waters interact rapidly with minerals in the rocks and dissolve certain chemical species, and when the water cools, minerals may be deposited.

Sources of the heat

Measurements in deep drillholes, throughout the world, show that the temperature of rocks within the Earth generally increases with depth; the rate

Temp (°C)	Density (kg/m^3)		Dynamic Viscosity $\mu \times 10^6$ (Pa.s)	
	Water (liquid)	Steam (vapour)	Water	Steam
20	998	0.017	1005	-
50	988	0.083	550	-
100	958	0.600	282	12.3
150	917	2.550	180	13.9
200	865	7.860	134	15.7
250	799	19.300	109	17.5
300	712	46.200	91	19.7

Table 9.1 Change in density and viscosity of saturated water and steam with temperature. Note the large changes with temperature. Data from Grant et al (1982).

Type of Water	pH	Cl (ppm)	Mg (ppm)	Na (ppm)	Li (ppm)	As (ppm)	B (ppm)	SiO_2 (ppm)
Freshwater	6-7	2-15	4	3-8	<0.1	<0.01		
Low temperature Water (1)	6.1	160	5	390	2	-	6.0	150
High temperature geothermal water (2)	8.2	2000	0.01	1200	10-12	4	25	700
Average seawater	7.8-8.3	19000	1300	11000	0.2	<0.01	4.6	5

(1) Copeland Springs, Westland (Barnes et al, 1978).
(2) Wairakei bore 27, a typical deep geothermal well.

Table 9.2 Chemistry of various types of waters

of increase, called the *geothermal gradient*, is 10 to 50°C/km, depending on the type of rock present and its geological history. This increase in temperature results from the conductive flow of heat from the interior to the surface; about 80% of this heat comes from radioactive decay of isotopes in minerals in the crust, and about 20% from primordial heat associated with formation of the Earth.

An additional, but localised, source of heat is from volcanic activity associated with plate tectonics. The skin of the Earth consists of numerous, thin, relatively rigid *lithospheric plates* which are in constant motion, moving at rates of 1 to 5 cm/yr relative to each other. In some places (such as the North Island) the edge of one plate (Pacific Plate) is pushed down beneath another (Australian - Indian Plate). Where this occurs, the crustal rocks at the edge of one plate are pushed down to depths of hundreds of kilometres where they melt (800 - 1300°C). This molten rock rises in stages towards the surface as a *magma body*; and may eventually reach the surface to form volcanoes (Stevens, 1985).

Types of Geothermal Systems

New Zealand's geothermal waters can generally be separated, on the basis of their temperature, geological location, and chemistry, into two main groups:

Low-temperature Waters with temperatures of about 20°C (ambient) to about 100°C, associated with active faults.

High-temperature Waters with temperatures greater than 100°C, associated with areas of active volcanism.

These waters form convective, hydrological systems within the upper part of the crust which are driven by the heat sources.

Low Temperature - Tectonic Geothermal Systems

Origin and Distribution

Isotopic analysis indicates that the water in low-temperature geothermal springs is mainly *meteoric* (rainfall) in origin. Rainwater slowly percolates down from the surface to depths of up to 5 km where it is heated by the warmer rocks. The warm, buoyant water then rises rapidly, along paths of high permeability such as faults, to emerge at the surface as hot springs or seeps, or mix with the groundwater over a small area. The presence of only very small amounts of tritium, a radioactive isotope of hydrogen, in such spring waters indicates that the *circulation time,* from rainfall to emergence of the water, is usually greater than 25 years.

In New Zealand, low-temperature springs occur mainly in the northern half of the North Island (Figure 9.1) and in a band through the central part of the South Island (Figure 9.2). In the North

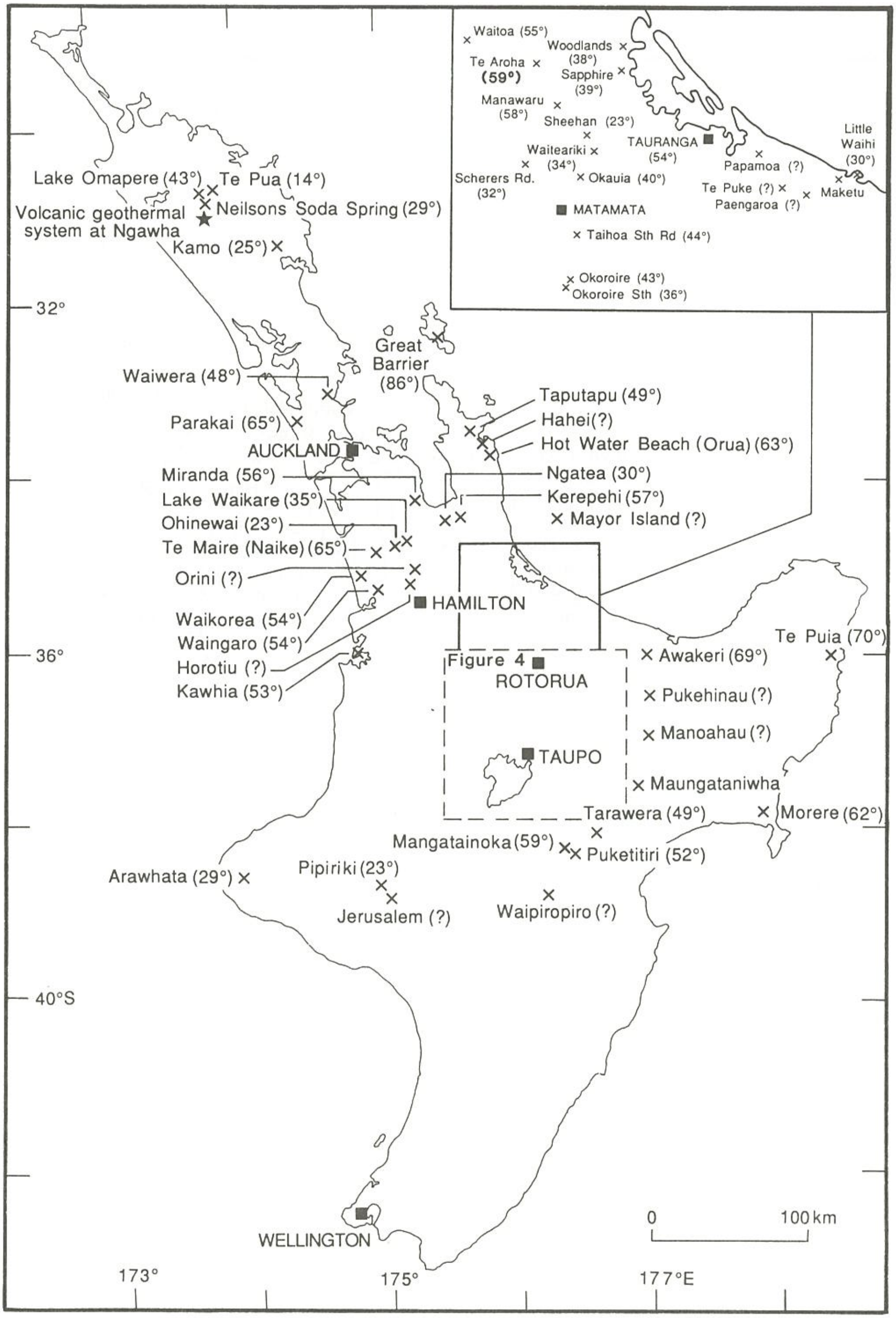

Figure 9.1 Map of North Island showing the location and maximum temperature (°C) of low-temperature waters (after Mongillo and Clelland, 1984).

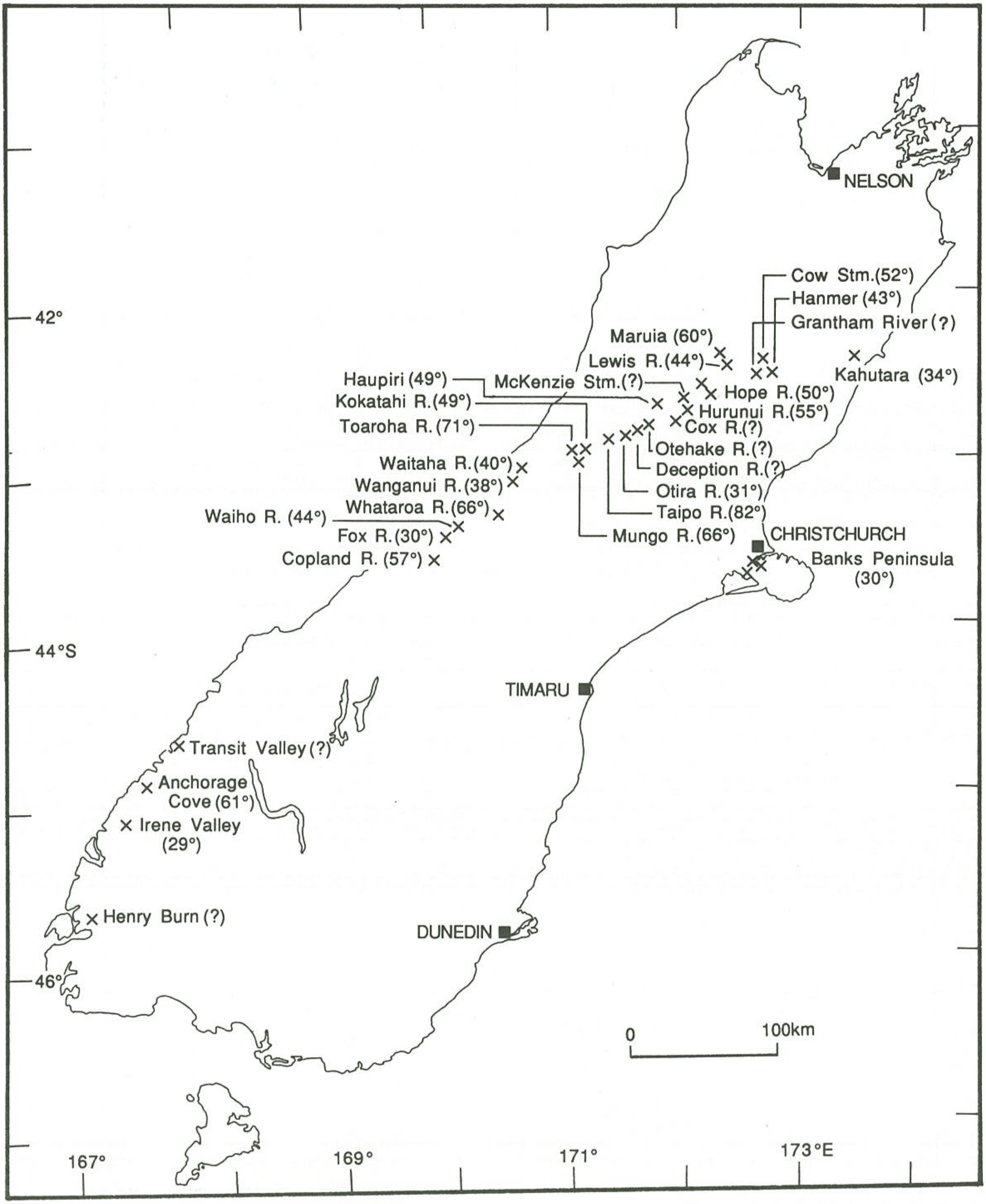

Figure 9.2 Map of South Island showing the location and maximum temperature (°C) of low-temperature waters, (after Mongillo and Clelland 1984).

Island, low-temperature springs are found mainly in areas of recent tectonic activity (such as the East Coast), and in areas of extinct volcanism (such as the Waikato and Coromandel). It might be expected that there would be numerous low-temperature springs in the south-eastern part of the North Island, where active faulting is common, but this is not the case. In the South Island, low-temperature springs are found mainly near the Alpine and Hope Faults, and appear to be associated with greater than normal temperatures (up to 200°C) at shallow depth (<2 km) which result from recent tectonic uplift (Figure 9.3) in these areas (Allis et al, 1979).

Chemistry

The chemical composition of tectonic geothermal waters varies widely (Table 9.3). The composition is influenced mainly by the chemistry of the rocks the water has come into contact with, and the temperatures to which the water has been raised. Indeed, the quantities of various forms of dissolved silica (SiO_2) can be used as a geothermometer to estimate the maximum temperature the water has reached. Similarly, the relative quantities of sodium (Na), potassium (K), and magnesium (Mg) are controlled by the maximum temperature reached and can also be used as a geothermometer. Despite the relatively low discharge temperatures of South Island springs (40 - 60°C), their chemistry indicates that some of the waters have reached temperatures of over 100°C (Allis et al, 1979). The fluids within rocks with which the water interacts will also influence the chemistry. For example, the waters from Te Puia and Morere Springs, on the East Coast of the North Island, are highly mineralised; the high concentrations of Na and Cl (Table 9.3) are probably the result of mixing with connate or fossil water from the underlying marine sediments.

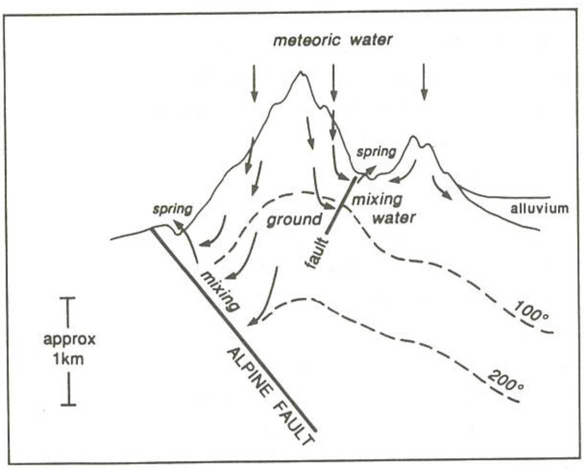

Figure 9.3 Scheme for the origin of low-temperature tectonic geothermal hot springs (adapted from Allis et al, 1979).

Exploration and Use

There has been no systematic exploration for low-temperature systems in New Zealand; those not already known to the Maoris before arrival of Europeans have generally been found accidentally by explorers, mineral prospectors, farmers, and trampers. It is unlikely that any new low-temperature geothermal springs of significant size or flow rate will be discovered, but it is probable that additional localised areas of warm groundwater will be found.

The main use of low-temperature waters in New Zealand is for bathing; public baths operate at Kamo, Waiwera, Parakai, Miranda, Te Maire, Waingaro, Hot Water Beach, Okauia (Matamata), Te Aroha, Tauranga, Awakeri, Te Puia, Morere, Hanmer, and Maruia. A history of the use of some of these springs is given by Rockel (1986).

High-Temperature Volcanic Geothermal Systems

Origin and Distribution

The fluids of volcanic geothermal systems are hotter, have distinct chemical composition, and are more restricted in occurrence, than those of tectonic systems. Meteoric waters percolate down from the surface to depths of 5 - 10 km where they become heated by hot (up to 800°C) volcanic rock,

Location (See Figs.1 & 2)	Temp (°C)	Flow (l/s)	pH	Cl	Na	K	Mg	SiO_2	SO_4	HCO_3
				(ppm)						
Kamo	24	0.2	6.9	230	220	14	*	102	10	1330
Waiwera	48	*	8.0	1032	635	7	2	38	4	24
Te Aroha	85	0.9	8.3	581	3162	40	4	125	388	7000
Okauia	40	1.5	*	25	179	13	16	127	9	*
Te Puia	59-70	*	7.2	7770	4400	34	25	48	70	175
Morere	62	*	6.7	15500	7100	100	135	28	21	175
Tarawera	38-49	*	8.4	637	475	5	1	40	19	225
Hanmer	53	8	8.4	476	388	5	0.1	46	25	187
Lewis River	44	0.02	8.2	10	109	4	0.2	46	3	279
Maruia	65	5	8.6	139	153	7	0.1	71	82	145
Wanganui River	40	5	6.9	57	109	6	0.6	41	6	227

Data from Petty (1972), Barnes et al (1978), Lawless et al (1981), R B Glover (personal communication).

TABLE 9.3 Chemical composition of selected tectonic geothermal waters. Note the wide variety in composition, especially of Cl.

such as a magma body. Although shallow magma bodies are believed to be the prime source of heat for high-temperature geothermal waters, no such body has yet been positively identified beneath any of the known volcanic geothermal systems in New Zealand, and their shape, size, depth, and chemistry remain largely speculative. The waters become heated and interact chemically with the volcanic rock then, being of lower density than the surrounding water, they rise towards the surface through pores and fractures in the overlying rocks.

At the surface, the volcanic geothermal systems are characterised by groups of thermal features (described later) within an area of 5 to 15 km^2. The natural discharge of energy from these features can be considerable: before exploitation, Wairakei Geothermal Area discharged more than 400 Megawatts (MW) of heat, with one feature alone (Karapiti Blowhole) discharging 12 MW (Allis, 1981).

High-temperature waters are associated only with active volcanism and in New Zealand are confined to the northern and central parts of the North Island (Figure 9.4). A list of the larger, known volcanic geothermal systems is given in Table 9.4.

Maori tradition has a different origin for the volcanic geothermal systems. Their legend (Grace, 1959) is that when Ngatoroirangi (chief of the Te Arawa tribe) and Ngauruhoe (his slave), were exploring the Taupo Volcanic Zone they climbed the slopes of Mount Tongariro and were close to dying from the cold. Ngatoroirangi called to his sisters, Kuiwai and Haungaroa, in Hawaiki across the Pacific Ocean, to send heat to resuscitate them. The sisters heard his cry for help and, with the fire gods Pupu and Te Hoata, set out underground to bring heat. In the search for their

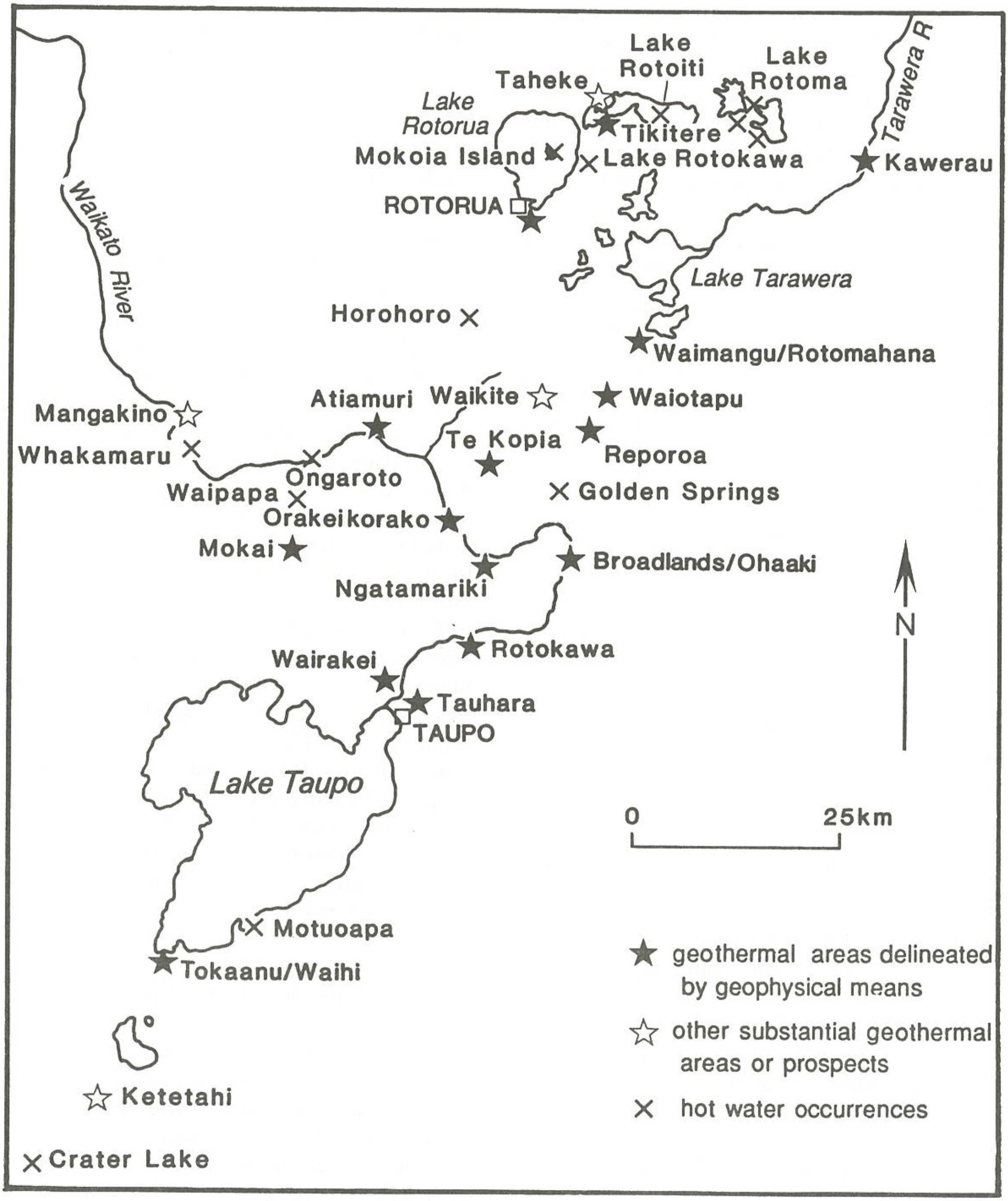

Figure 9.4 Map of the central part of North Island showing the location of known volcanic geothermal systems, (after Mongillo and Clelland, 1984).

brother they stopped and came to the surface to look for him at places on the way. Their route led from Whakaari (White Island), via Moutohora (Whale Island), Okakaru, Rotoehu, Rotoiti, Tarawera, Paeroa (Waikite), Orakeikorako, Taupo, and Tokaanu, to Tongariro. Where they stopped, the heat burst out as thermal activity, and remains as ngawha (overflowing pools), puia (volcanoes, hot springs), and wairiki (hot springs).

System	Approx Area (km^2)	Temp °C	Status	No. of Drill holes	Available Energy (PJ)
Horohoro	5	220	f	1	400
Ketetahi	?		u	-	?
Kawerau	10	250	e	32	1 300
Mangakino	5	220	f	1	400
Mökai	15	280	f	6	2 700
Ngatamariki	10	260	f	4	1 400
Ngawha	15	230	f	15	1 400
Ohaaki	10	260	e	46	1 400
Orakeikorako	5	260	p	4	?
Reporoa	5	230	u	1	500
Rotorua	5	220	u	-	400
Rotokawa	15	280	f	8	2 700
Taheke	5	230	u	-	500
Tauhara	15	240	f	5	1 900
Te Kopia	5	240	f	2	500
Tikitere	10	230	p	-	900
Tokaanu	10	250	u	-	1 300
Waikite	5	230	u	-	500
Waimangu	10		p	-	?
Wairakei	15	230	e	135	1 400
Waiotapu	10		p	7	?

Temperature: Inferred average temperature over 3 km depth range.

Status: p = protected for tourist use;
e = commercially exploited;
u = uninvestigated;
f = some feasibility studies made.

PJ = Petajoule = 10^{15} joules.

Table 9.4 New Zealand's volcanic geothermal systems; taken from Allis and Speden (1991) and Lawless et al (1981).

Conceptual Model

Only the upper parts of volcanic geothermal systems have been investigated by drilling; the deeper parts have been probed by geophysical techniques and their behaviour is deduced theoretically from mathematical models or from scale models in the laboratory. A conceptual model of a volcanic geothermal system (Figure 9. 5) is that of a slowly rising plume of hot water heated by some very hot (>500°C) body at depth (>5 km). Once established, this plume is remarkably stable and the water rises vertically, seeking permeable paths to the surface. Although the temperatures may be several hundred degrees, the pressures at depth are such that the water does not boil (Table 9.5). As the water rises, cooler water from the adjacent rocks may be entrained in the plume, reducing the temperature and diluting the concentrations of dissolved salts. Near the surface a number of factors can complicate the flow: reduced pressures can cause local boiling, the geological structures in the rocks can channel flow, and surface waters can infiltrate the system.

Earlier conceptual models visualised the plume to be shaped like a mushroom with the narrow stalk being the region of upflow and the much larger head being the geothermal system that is exploited (Elder, 1981). The water on the edges of the system was believed to cool and sink, mixing with deep inflowing meteoric water, to be heated again by the hot source body or the plume. However, more recently developed models favour a "once through" concept, with the cross-sectional area of the plume decreasing towards the surface (Figure 9.5).

Factors Controlling Fluid Movement

The most important factors controlling the movement of fluid in volcanic geothermal systems are pressure and temperature; differences in chemical composition of the water have little effect on the movement, except when large amounts of gas (mainly CO_2) are present.

Measurements in drillholes (Figure 9. 6) show that in most unexploited geothermal systems in New Zealand the pores and fractures in the rock

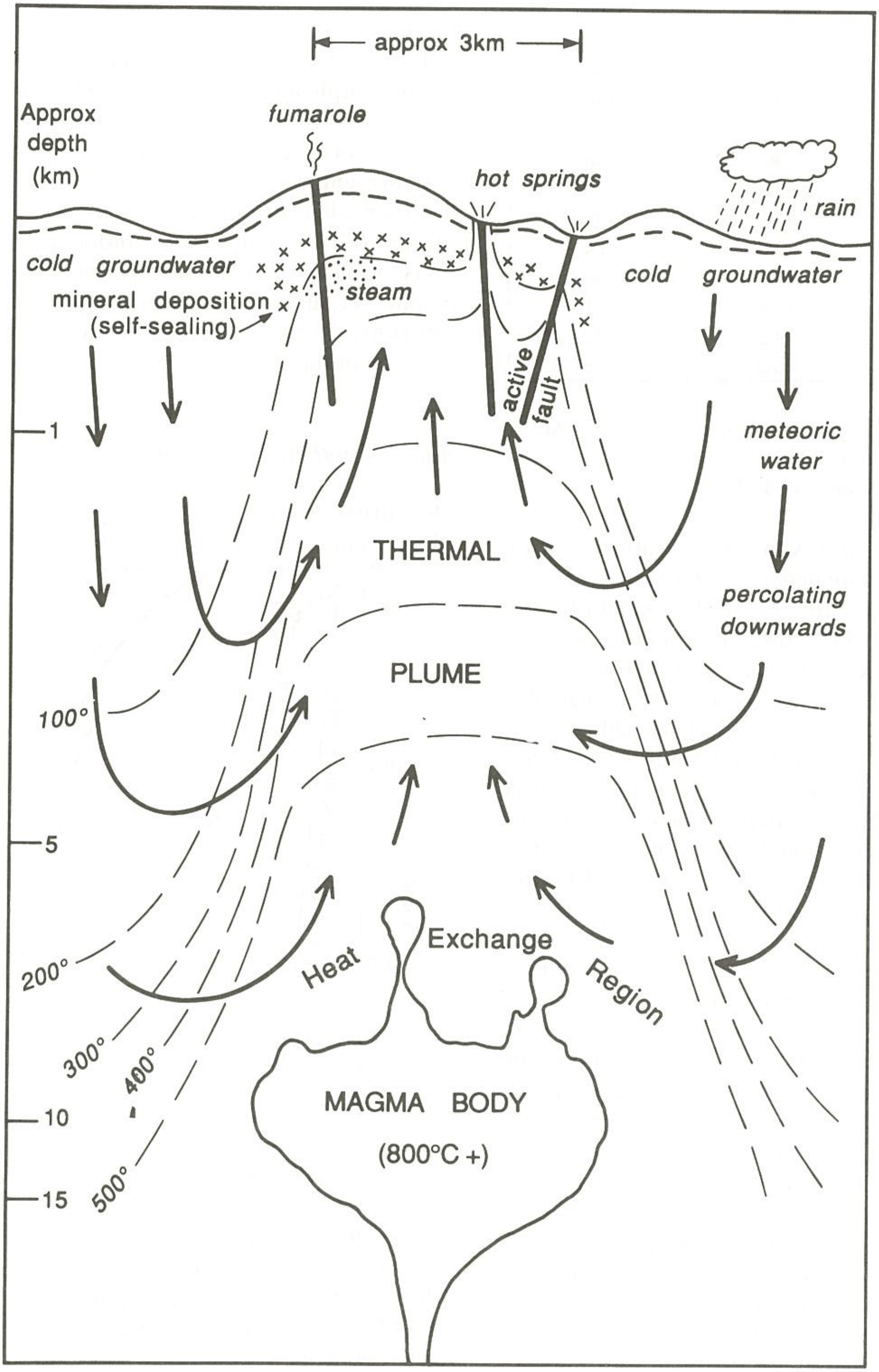

Figure 9.5 Conceptual model of a high-temperature, volcanic geothermal system. Scale is only approximate.

Depth (m)	Pressure (MPa)	Temperature at boiling (°C)
0	0.10	100
50	0.56	156
100	0.99	180
200	1.85	208
300	2.68	228
400	3.49	242
500	4.28	254
1000	8.00	295
1500	11.4	321
2000	14.5	339
3000	20.0	366

Table 9.5 Changes in the boiling point of water, with pressure and depth.
1 MPa = 10^6 N/m^2 = 10 bar.

are filled with water whose temperature is at or near *boiling point for depth*. As depth increases, so also does the pressure, due to the weight of overlying fluid, and this causes the boiling point to increase. The temperatures at which water will boil, assuming the pressure at depth is that of a column of boiling water, are given in Table 9.5 and shown in Figure 9. 6. If the hot water in the plume rises fast enough, the boiling point for depth is exceeded and the water boils. The resultant, very buoyant steam rises much more quickly towards the surface, where it is either condensed by the cold overlying groundwater or escapes as a fumarole.

The *permeability* of the rocks controls the rate of movement of water in a geothermal system. In a cold groundwater system the *horizontal* permeability of the aquifer is the most important parameter, but in a geothermal system (except very close to the surface) it is the *vertical* permeability which matters. In most geothermal systems the permeability of the fractures is more important than that of the pores in the rock. A representative sample of a geothermal system may therefore be tens or hundreds of metres in size, and it is not possible to make meaningful measurements of permeability in the laboratory, as it is for cold groundwater systems. Estimates for the vertical permeability at Wairakei, determined from well testing and mathematical modelling (Grant et al, 1982), range from 10 to 50 millidarcy (md) (1 - 5 x 10^{-14} m^2).

Fluid movement in volcanic geothermal systems is also complicated (particularly in the upper parts of the system and during exploitation) by the presence of steam. If two *phases* -water and steam - are present, each reduces the flow of the other to less than what it would be if only one phase were present (Grant et al, 1982). Under certain conditions one phase may be mobile and the other immobile. To further complicate matters, one phase can convert to the other as the fluid moves from one part of the system to another.

Age of Geothermal Systems

Measurements of heavy isotopes of hydrogen (deuterium and tritium) present in geothermal

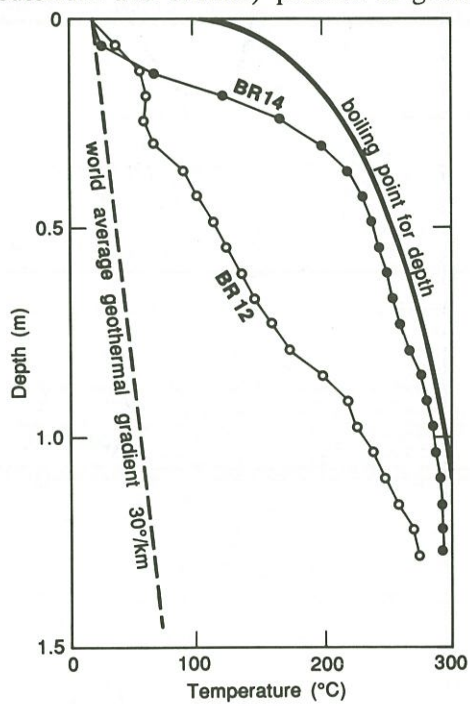

Figure 9.6 Graph showing the variation of temperature with depth in two drillholes in the Ohaaki (Broadlands) Geothermal Field. Well BR14 lies near the centre of the field and for most of its depth the temperature lies close to boiling point for depth curve; Well BR12 lies on the edge of the field. Data from Smith (1970).

waters suggest that the circulation time for high-temperature geothermal waters in New Zealand is longer than 100 years, but less than 12 000 years (Stewart, 1978). Cooling of the source body eventually causes the system to decay and die, and geological evidence suggests that individual systems may exist for up to half a million years (Grindley, 1965; Browne, 1979). Epithermal mineral deposits, such as at Ohakuri, are the fossil remnants of previously active volcanic geothermal systems.

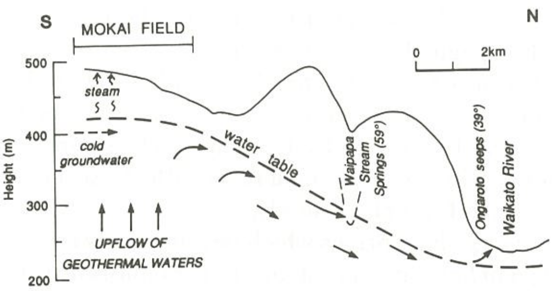

Figure 9.7 A model explaining the origin of the Ongaroto Seeps and Waipapa Stream Springs, associated with an outflow from the Mokai Geothermal Field (adapted from Bibby et al, 1984).

Relation to Groundwater

When upwelling hot water in a high-temperature system nears the surface it meets large volumes of overlying cold groundwater. The amount of hot water reaching the surface depends on the vertical permeability beneath that area. In many places the permeability is reduced by *self-sealing* resulting from silica (SiO_2) precipitating out of the geothermal water when it cools, and forming a zone of reduced permeability or "cap" to the system. This explains, in part, why the surface features of a volcanic geothermal system are often restricted to only a small part of the area of the system at depth.

Hot geothermal fluid may mix with and warm the cold groundwater in areas of high vertical permeability. If this occurs in a topographically high area and the groundwater is flowing naturally towards a topographic low, the resultant warm water or *outflow* may be carried a considerable distance from the area of the geothermal system. At Mokai (Figure 9.7), hot geothermal waters rising from Mokai geothermal system meet and mix with the groundwater flowing down slope, and the resulting warm water is carried laterally about 10 km to emerge in the bed of the Waikato River as the Ongaroto Springs (39°C). Outflows are sometimes very large and hot, and have been mistakenly thought to indicate the presence of a geothermal system below the place where they emerge, until drilling or geophysical measurements show that only a thin layer of hot water is flowing near the top of a zone of cold groundwater.

In places where good vertical permeability occurs, such as where the near surface rocks are broken by active faults, hot fluid can rise readily to the surface, producing characteristic discharge features. The type of feature and its size is controlled by the amount of hot geothermal water or steam reaching the surface. Thermal features change naturally with time, not only in type and output, but also in location, area, and chemistry as a result of silica precipitating in the fractures and blocking the upflow, the opening of new fracture paths by fault movement, climatic changes, and variations in groundwater level and temperature. It is often difficult, however, to clearly separate the effects of natural changes from those induced by exploitation.

Natural Discharge Features

Most volcanic geothermal systems have distinctive surface features associated with the escape of geothermal water or steam at the surface. However, the area, temperature, or flow rate of the surface features is no indication of the size of the system at depth; this can be estimated only by geophysical and geochemical tests, and ultimately determined by drilling. Common surface features of these systems are:

- ☐ *Hot springs*, with flow rates ranging from less than 1 to more than 100 l/s (Frying Pan Lake at Waimangu).

- ☐ *Hot pools*, with temperatures up to boiling point, and covering areas of 1 to 100 m^2 (e.g. Champagne Pool at Waiotapu).
- ☐ *Mud pots ("porridge pots")* - small, often ebullient, hot pools containing mud, formed when near surface groundwater is heated by steam to form acidic waters which react with the surface rocks to form clays (mud).
- ☐ *Geysers* - hot springs which intermittently erupt a jet of hot water and steam. (e.g. Pohutu Geyser at Whakarewarewa (Rotorua), which erupts to a height of about 35 m).
- ☐ *Fumaroles ("blow-holes", "solfataras")* - vents from which steam is emitted under pressure.
- ☐ *Thermal ground* - areas of warm, hot or steaming ground, often with little or no vegetation because the heat has killed, or stunted, plant growth.
- ☐ *Areas of sinter* - a hard siliceous (or sometimes calcareous) precipitate formed when silica-saturated geothermal water from a spring or geyser cools as it flows across surrounding areas of flat ground.
- ☐ *Phreatic explosion pits* - formed when boiling liquid groundwater, lying close to the ground surface, suddenly flashes to steam and the overlying material is thrown into the air by the force of the sudden eruption (e.g. Karapiti Thermal Area at Wairakei).

Descriptions and excellent photographs of geothermal features are given in Lloyd (1972, 1975), and historical accounts of the use of the waters for medicinal purposes are given by Herbert (1921) and Rockel (1986).

Chemistry

The chemical composition of waters associated with volcanic geothermal systems is the result of the interaction of meteoric water with hot rocks in the upper part of the Earth's crust; juvenile and magmatic water components from the heat source are negligible.

Water samples taken from deep drillholes contain high concentrations of sodium and potassium chlorides, and silica (SiO_2), and significant concentrations of sulphate (SO_4), bicarbonate (HCO_3), fluoride, lithium, and boron (Table 9.6). These waters are called *alkali chloride waters*, and represent the water in the deep part of the plume (Figure 9. 5).

As the plume nears the surface, the reduction in pressure may cause boiling and the formation of steam. The steam contains a large proportion of the gases previously dissolved in the water. The most common gases are carbon dioxide (CO_2) and hydrogen sulphide (H_2S), together with minor amounts of hydrocarbons, hydrogen (H_2), and ammonia (NH_3) (Table 9.7). The concentration of gases in the steam is an important factor in the design of geothermal power generation plants.

Near the surface, waters of different chemical composition often occur. When the steam condenses into the groundwater the H_2S in the steam is oxidised to sulphate and *acid sulphate waters* are formed. These waters generally have low chloride content, low pH, and the constituents are leached mainly from rocks surrounding the discharge feature (Table 9.6). Low chloride, *bicarbonate waters* form when steam containing CO_2 and H_2S condenses into an aquifer. Under certain conditions, reaction with the rock produces neutral-pH bicarbonate or bicarbonate-sulphate solutions with high Na concentrations (Table 9.6). Another type of water found is a *acid sulphate - chloride water*, which is formed when alkali chloride and acid sulphate waters become mixed. The water types are identified from the concentrations of critical chemical species: Cl, Na, HCO_3, and SO_4. The effects of dilution or mixing are recognised by comparing the concentrations with the molecular ratios of specific species; e.g. Cl/SO_4.

Exploration Techniques

Volcanic geothermal systems contain much more thermal energy than tectonic systems, and have been actively sought. Exploration methods for volcanic geothermal systems, developed largely in New Zealand, use the fact that most (but not all) hot geothermal waters contain relatively high concentrations of dissolved salts (Table 9.2) and are thus very good conductors of electricity; i.e. they

Water Type	pH	Cl	Na	B	Li	F	HCO_3	SiO_2	SO_4	Cl/SO_4 (mr)	Cl/B (mr)
		(ppm)									
Alkali Chloride (1)	8.3	1743	1050	48.2	11.7	-	-	805	8	602	11.0
Acid Sulphate (2)	2.8	32	11	2.5	-	-	0	280	347	0.25	3.9
Bicarbonate (3)	8.6	2.7	230	0.5	1.2	3.7	680	191	11	0.68	1.6
Acid Sulphate - Chloride (4)	2.8	612	405	10.1	-	-	0	370	666	2.5	18.5

mr Molecular ratio
(1) Drillhole BR2, 1030 m depth, Ohaaki Geothermal Field
(2) Hot pool at Waiotapu
(3) Drillhole WK5, 471 m, Wairakei Geothermal Field
(4) Hot pool at Waiotapu

Table 9.6 Examples showing the different chemistry of the various water types. Data from Ellis & Mahon (1977)

have low electrical resistivity. The location and extent of geothermal waters can thus be found from electrical resistivity measurements.

Reconnaissance prospecting often begins with a resistivity survey. An electrical current (generally 200-400 V,d.c.; and 0.1-1 A) is passed through the ground between two current electrodes set 1 or 2 km apart, and the electrical response (usually 0.01-1 mV) is measured between a pair of potential electrodes, 50 or 100 m apart, situated halfway between the current electrodes. The *apparent resistivity* of the ground between the current electrodes is then calculated using Ohms Law (Bibby, 1988). The electrode array is moved from place to place, the apparent resistivities at each site determined, and the values contoured to produce a resistivity anomaly map (Figure 9.8). Rocks at shallow depths (<1 km) containing hot geothermal waters of a volcanic geothermal system show up in such anomaly maps as areas of low resistivity (typically less than 10 Ωm), whereas rocks outside the system have high resistivity (100 - 10 000 Ωm).

Source	Source Depth (m)	Temp (°C)	Steam Fraction	Gas content (mole %)	Gas Composition (mole %)				
					CO_2	H_2S	Hydro-carbons	H_2	NH_3
Fumarole, Wairakei	0	115	1.00	0.17	94.6	2.3	0.7	1.0	0.3
Average well at Wairakei	650	260	0.32	0.06	91.7	4.4	0.9	0.8	0.6
Well BR11, Ohaaki	760	260	0.36	0.61	94.8	2.1	1.2	0.2	0.2

Table 9.7 Composition of steam (at 1 bar pressure) in fumaroles and wells; data from Ellis and Mahon (1977).

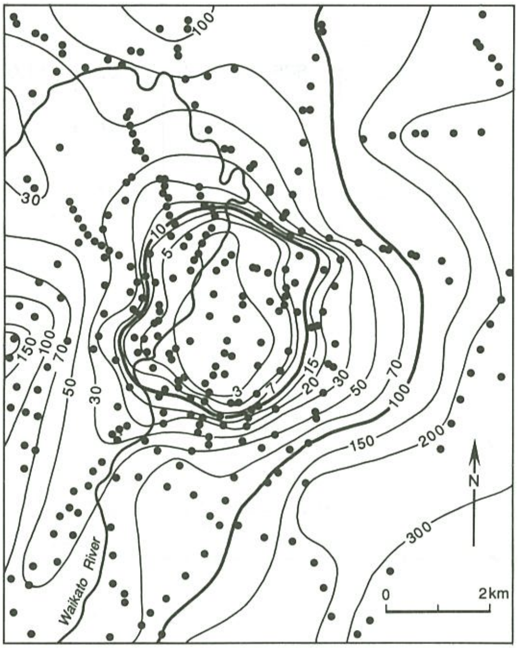

Figure 9.8 Electrical resistivity anomaly map of the Ohaaki (Broadlands) area, showing the low resistivity region associated with the high-temperature waters. Contour lines show apparent resistivity in Ωm; solid dots indicate the location of measurement points using a current electrode separation of 1 km.

Not all areas of low resistivity are associated with volcanic geothermal systems; extensive areas of clay mineral alteration, and sea water intrusions near the coastline may also produce a similar resistivity "signature". It is therefore necessary to combine, or follow up, the resistivity surveys with geochemical, geological, and other geophysical measurements to confirm the presence of a volcanic geothermal system. When the presence of a system has been confirmed, more detailed electrical resistivity surveys are conducted to precisely locate the boundaries of the hot water. Several deep exploration wells may then be drilled to verify the scientific data and to test if there is sufficient permeability in the rocks to allow continuous withdrawal of the hot water.

Exploitation

At present the greatest use of geothermal water in New Zealand is in the generation of electricity. When a deep geothermal well is discharged, the emerging fluid is a mixture of liquid water (about 80% by mass) and steam. The steam is separated and piped to the power station. The water, together with condensed steam from the power station, is disposed of, either by putting it into a nearby river or by pumping it back into the ground (*reinjection*).

Exploitation of a geothermal system can have profound and complex effects, both on the system and on the overlying cold groundwater; the changes induced are only just being recognised and understood. Wairakei is the only geothermal system in New Zealand which has been exploited sufficiently, and over a long enough time period, to show these effects. Other systems may behave differently; time will tell.

Effects of exploitation

1. Pressure Changes

The initial effect of exploitation of the Wairakei geothermal system was a reduction of pressure near the top of the system which caused the formation and expansion of a *two-phase zone* containing both steam and water. This zone has two parts: an upper *steam-zone* in which steam is the continuous pressure-controlling phase, and a lower *liquid-dominated zone* in which water is the continuous phase but some steam may be present. During the early stages of exploitation, the two-phase zone expanded laterally towards the boundaries of the system. Continued exploitation caused the two-phase zone to expand vertically downwards and pressures in the underlying *deep-liquid zone* (single-phase) to decrease. This process is seen in the changes with time in both pressure and temperature (Figures 9.9 and 9.10), and is shown schematically in Figure 9.11.

By 1972, after 15 years of exploitation, the extraction of 867 Mt of fluid had resulted in deep-liquid pressures at Wairakei falling by 2.5 MPa (25 bar) (Figure 9.9), but as exploitation

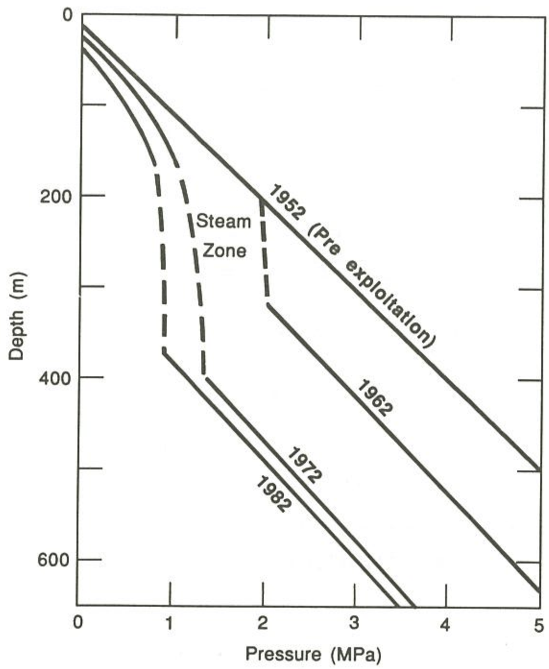

Figure 9.9 Graph showing the variation of pressure with depth, and changes with time, in the eastern part of the borefield at Wairakei. Data from Allis and Hunt (1986). Broken line represents the steam zone. Note the change in deep pressures with time as the steam zone formed and expanded vertically.

continued the rate of vertical expansion of the two-phase zone decreased and stabilised. At most places in Wairakei, the deep-liquid pressure has been near constant from 1970 to 1990 and the two-phase zone is now about 300 m thick. However, in some places the deep-liquid pressures began to increase in the early 1980's, indicating that here the two-phase zone was becoming thinner, due to inflows of water from the overlying cold groundwater zone. In most other places, saturation in the steam zone has decreased: in other words the amount of liquid water in the pores and fractures of the rock has decreased.

2. Temperature Changes

Temperature changes are much more difficult to measure because they require a well to be shut-off and left standing for several months for the temperature inside to equilibrate to that of the rock outside. Interpretation can also be difficult because of internal flows (up or down the well) between different horizons, even after by the well has been shut off. In most places at Wairakei, deep-liquid temperatures fell by about 20°C during the first 10 years of exploitation, but have since stabilised at about 25°C below the pre-exploitation value (Figure 9.12a). In some other places, however, temperatures in the two-phase zone have decreased by up to 100°C as a result of cold water invasion (Figure 9.12b).

3. Mass Changes

As a result of exploitation at Wairakei, nearly 2 km^3 of water, weighing about 2 billion kilogrammes, have been withdrawn and disposed of into the Waikato River and thus lost from the system. Microgravity measurements at Wairakei have shown that in the early stages of exploitation, during formation and expansion of the two-phase zone, fluid was removed and only a small propor-

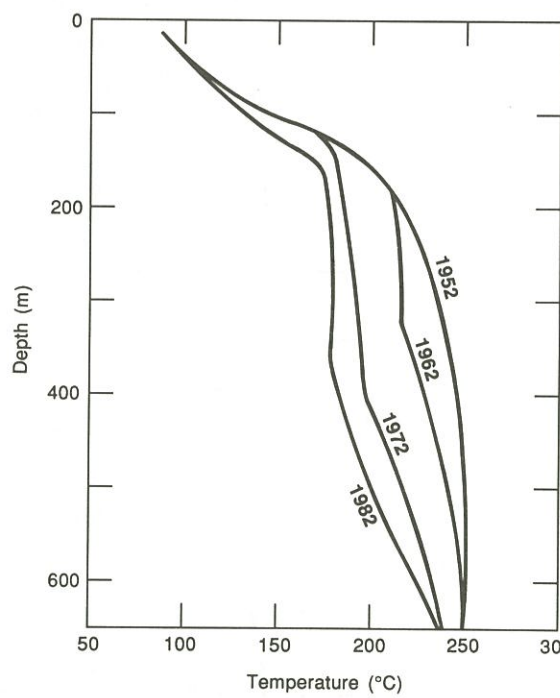

Figure 9.10 Graph showing the changes in average temperature with depth, and with time, in the eastern part of the borefield at Wairakei (after Allis and Hunt, 1986).

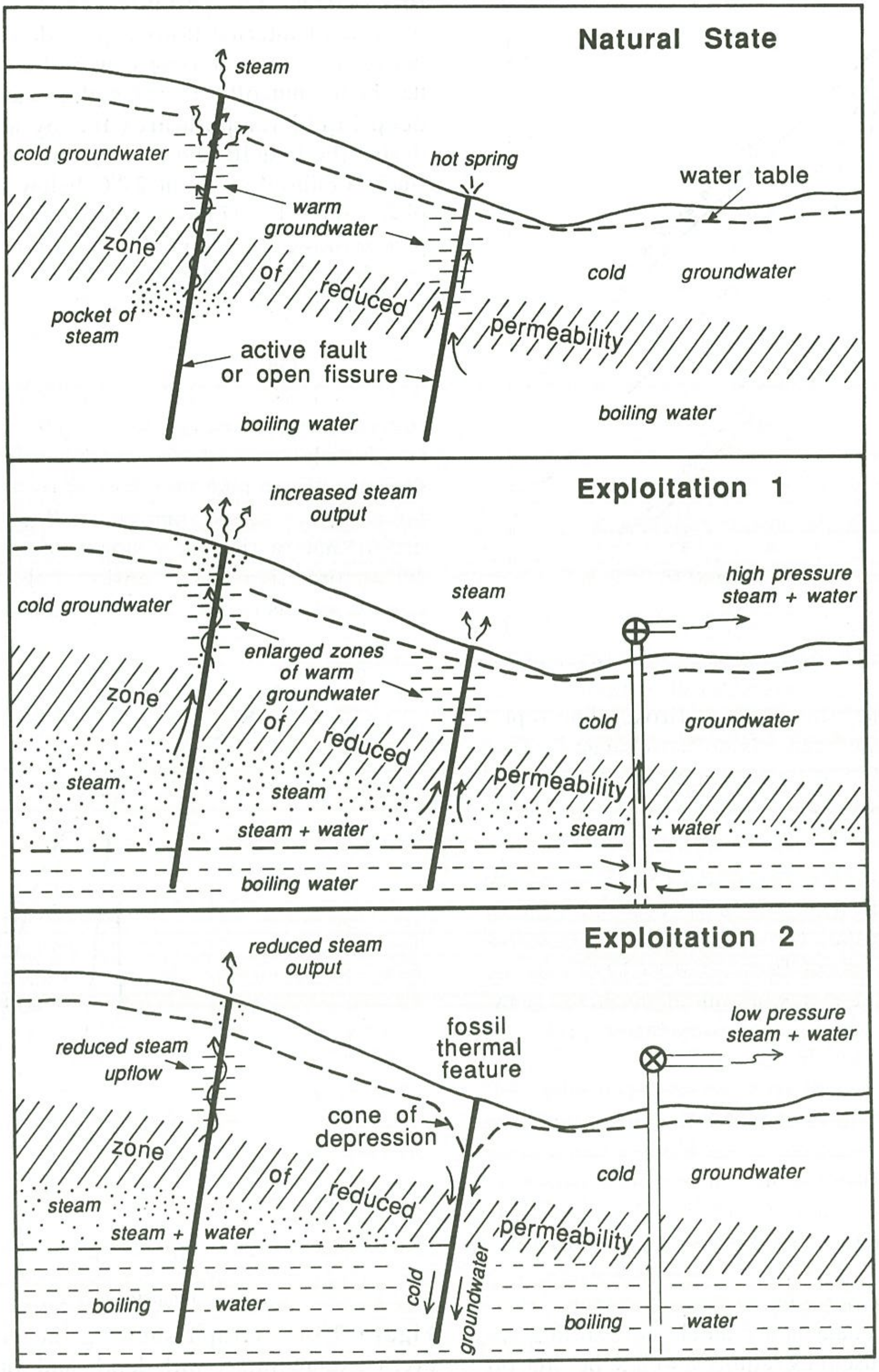

Figure 9.11 Diagrams showing changes that can occur in the upper part of a volcanic geothermal system during exploitation.

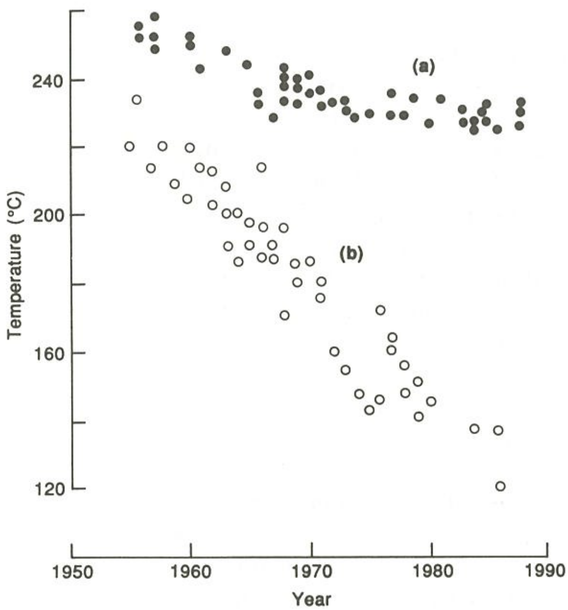

Figure 9.12 Graph showing feed temperatures of (a) deep-liquid fed wells, and (b) wells in two-phase zone affected by cold water invasion, at Wairakei. Data from Electricorp (1990).

tion was replaced; the geothermal system was in effect being mined. However, from the 1970's onwards recharge increased, and total inflow of water (both hot and cold) now roughly balances outflow (Allis and Hunt, 1986).

4. Changes in Surface Heat Flow and Thermal Activity

The formation and expansion of a two-phase zone in the early stages of exploitation alters heat flow (Figure 9.11). Steam is much more mobile than water; it can move through small fractures impervious to water and can move much more quickly through larger fractures.

At Wairakei, heat flow from natural discharge features was about 400 MW prior to the start of exploitation in 1958, increased to a peak of nearly 800 MW by the mid 1960's, and has since declined to about 600 MW (Allis, 1981). Most of this increase was associated with increased thermal activity in the Karapiti Thermal Area, 3 km south-west of the main production borefield. These changes have been attributed to steam from the newly formed steam zone rising to the surface along fissures that were previously impervious to water. At the same time, there was a rapid decrease in geyser and hot spring activity in Geyser Valley, 1 km north of the borefield; geyser eruption periods became longer, more irregular, and finally ceased because the upflow of hot water, essential for geyser activity, was replaced by steam.

5. Changes in Groundwater Level

As pressures in the steam zone fall due to production from the drillholes, the amount of steam and hot water passing upwards into the overlying cold groundwater decreases. If steam zone pressures fall sufficiently then the upward flow may cease, and hot springs, seeps, and other surface features will die. If pressures fall further, then the flow may reverse and cold groundwater will flow down into the geothermal system (Figure 9.11); the low permeability "cap" now becomes an "umbrella", albeit a leaky one, protecting the top of the system from being quenched by the cold groundwater. At Wairakei, in one small (1 km^2) part of the borefield the groundwater level has fallen by more than 30 m, mostly since the early 1970's, forming a cone of depression in the surface of the groundwater.

6. Ground Subsidence

Exploitation can also lead to localised ground subsidence. Levelling surveys have shown that subsidence is occurring at all three commercially exploited geothermal fields in New Zealand (Wairakei, Ohaaki-Broadlands, and Kawerau). At Wairakei, subsidence of up to 11 m has occurred since production began; this subsidence is not centred in the area of maximum production but in a 1 km^2 area about 500 m north-east of the main borefield (Figure 9.13). The simplest explanation is that beneath this area some of the rocks above the system have high compressibility and have compacted, due to a gradual decline in fluid pressure, as the steam and liquid pressures in the underlying two-phase zone have fallen (Allis, 1990). The subsidence has led to deformation of casing in wells near the area (Bixley and Hattersley, 1983),

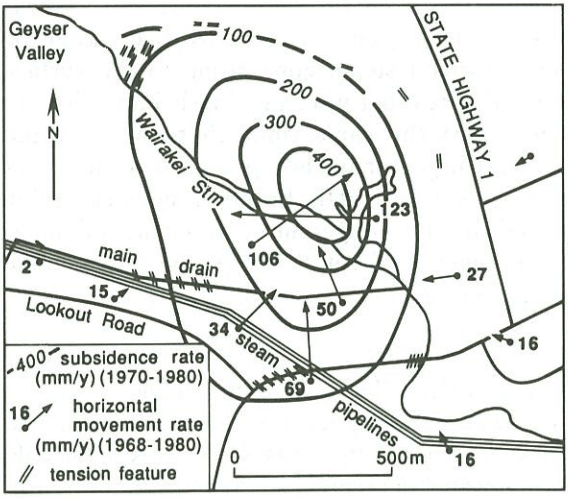

Figure 9.13 Map showing rates of ground subsidence and horizontal movement at Wairakei (adapted from Allis, 1990).

and has necessitated the mounting of steam pipelines on roller supports. The subsidence has also led to horizontal ground movements of up to 120 mm/year, tension cracks appearing on the ground surface, and ponding of the Wairakei Stream as a result of tilting of the stream bed (Allis, 1990).

Environmental effects

A 100 MW geothermal power station, tapping a liquid-dominated, volcanic geothermal system, produces about 100 million tonnes of waste fluid each year, of which more than 70% is hot saline water containing appreciable amounts of the elements arsenic, boron, fluorine, lithium, and mercury which in high concentrations are poisonous. The environmental effects of exploitation depend largely on the method of waste fluid disposal. Presently, all the liquid (both separated water and condensed steam) produced at Wairakei is cooled and discharged into the nearby Waikato River where it becomes diluted by a factor of about 100. Cooling can cause precipitates of antimony and arsenic sulphides to form which also contain metals such as mercury and thallium; these heavy precipitates then accumulate in the river bottom sediments (Ellis and Mahon, 1977). At Ohaaki (Broadlands) the environmental problems are minimised by reinjecting waste fluids deep into the edges of the geothermal system.

A 100 MW geothermal power station also releases into the atmosphere up to one hundred and fifty thousand tonnes of carbon dioxide and several thousand tonnes of hydrogen sulphide each year. However, by comparison, Wairakei geothermal power station produces less than 10% of the carbon dioxide that would be emitted from a coal-fired power station of the same capacity.

Reinjection

At present, reinjection is the most environmentally responsible way of disposing of waste fluid from schemes exploiting a volcanic geothermal system, but it is costly and is not a simple exercise. Firstly, rocks with high permeability which can readily absorb the waste fluid must be located. Secondly, there must be little or no hydraulic connection between these rocks and those of the production zone of the system, otherwise the cooler reinjected fluid will quickly return to the production wells. Thirdly, the reinjection system must be operated so as to avoid precipitation of minerals , mainly silica, in the reinjection wells or in the rocks near the wells, which reduces the flow rate in the wells, or the permeability of the rocks. To avoid precipitation the temperature and pressure of the waste fluid must be kept at or above silica saturation conditions. Ideally, reinjection is done into a deep, highly porous rock unit outside the geothermal system but this is usually uneconomic because of the cost of locating such rocks, drilling the holes, and transporting the waste fluid. In most geothermal development schemes, waste water is reinjected into wells just inside the edge of the plume.

At Ohaaki geothermal power scheme, most of the waste fluid (separated water and condensate) from the 24 production wells is reinjected, at a total rate of about 1000 t/hr, into 8 wells distant from the production area. The waste fluid has a temperature of 80-150°C and is pumped into the reinjection wells at well head pressures of 2 to 3.5 MPa. Large scale production and reinjection at Ohaaki began in 1989, and so far there has been no significant returns of reinjected water into production wells or loss of permeability through silica precipitation.

Waste fluid from small scale exploitation of low-temperature systems is usually discharged into nearby streams or shallow wells into the groundwater, and is rarely a problem because it does not contain large amounts of dissolved minerals.

Future Use Of Resources

Low-temperature waters have been widely used in New Zealand for several centuries and it is unlikely that any significant new sources will be found or that exploitation will greatly increase above the present figure of less than 3 Petajoules (PJ) (Allis and Speden, 1991).

High-temperature waters, however, have not yet been fully explored, and even the large, known, sources have barely been touched. Current estimates (Table 9.4) are that about 20 000 PJ of thermal energy is available in central North Island alone; only about 75 PJ is being used (Allis and Speden, 1991). By comparison, the total energy content of our coal resources (excluding lignite) is 37 000 PJ and gas/oil/condensate resources is 5 000 PJ. High-temperature geothermal systems therefore represent a major, indigenous, energy resource which, with proper exploitation management, could last 200 to 300 years (Allis and Speden, 1991).

Acknowledgements

We thank Dr Richard Glover and Tom Lumb for constructive criticism of the manuscript, and Carolyn Hume for preparation of the figures.

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10
Groundwater - The Hidden Resource

Hugh Thorpe

Why Is Groundwater Important?

Because it comprises about 91% of the liquid fresh water on earth.

Because in many parts of the world it is often the only water available.

Because it is stored in aquifers and less susceptible to variation in supply, quality or to pollution.

Because ground water can be drawn on during dry periods when surface waters are inadequate.

What Is An Aquifer?

An aquifer is a geological formation which holds water, and which will release this water at a rate fast enough to be useful. To hold water a geological formation must either be porous i.e. contain pores, or fractured. This property is called porosity, and if the pores are full of water the formation is saturated. For water to be able to move through the formation the pores must be interconnected, as in sand or gravel, or linked by fractures, as in hard rock aquifers. If the pores are microscopic as in a clay, water can move only very slowly, therefore clay formations are not aquifers. Pores can range from the just visible, for example in sand aquifers, to the cavernous channels in some limestone areas such as Waitomo or Takaka (See chapter 11).

Water, Porous Media and Types of Aquifers.

There is a distinction to be made between porous material containing water and containing a water resource. All porous materials below the water table (Figure 10.1) will be saturated and may contain large volumes of water, but it is only a resource when it can be extracted. Clays contain perhaps 50% by volume of water, but some of this is attached to the clay particles and the rest can move only very slowly through the exceedingly fine pores. Thus clays do not contain a useful water resource whereas sand, containing less water, say 30%, but with coarser pores, does.

Water usually enters an aquifer at a point by falling as precipitation and percolating down through the soil. At some depth the formation is saturated. If a well is sunk in such an area the level at which water naturally stands in the well is called the water table, and this approximately divides the higher unsaturated zone from the lower saturated zone. Between the two is a capillary fringe which may be up to a few metres thick and contains variable amounts of water under slight negative pressure. (Figure 10.2)

Water in an aquifer may flow away from the area where it fell as precipitation. If the flow through the aquifer occurs easily the formation is said to permeable and conversely if the water cannot flow the rock is impermeable. Fractured rock, gravels

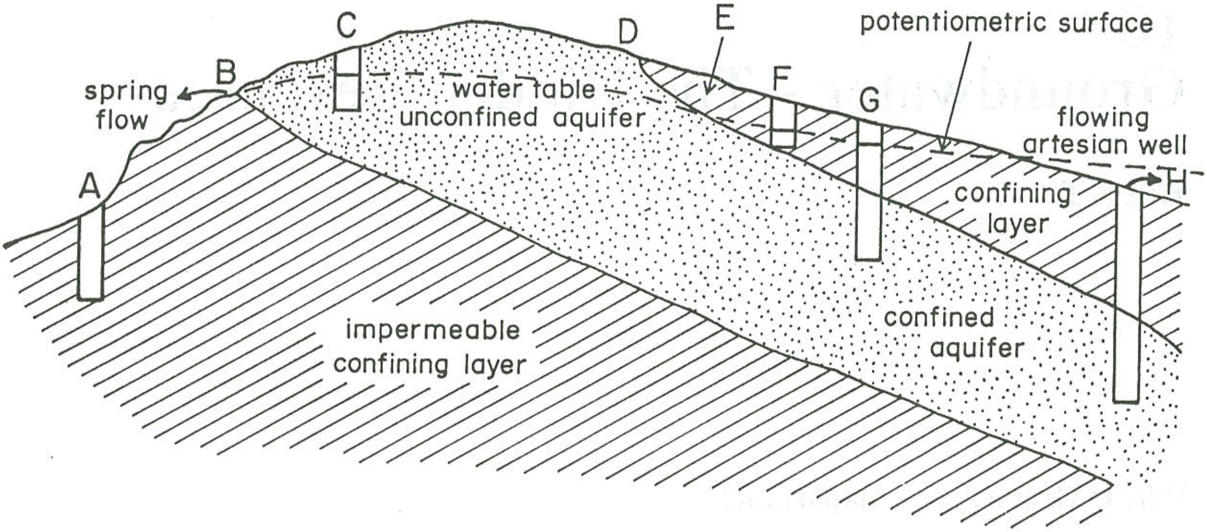

Figure 10.1 Aquifers. Between B and E the aquifer is unconfined and has a water table. To the right of E the aquifer is confined and has a potentiometric surface. Wells C, G and H would produce water with H flowing artesian. Wells A and F would produce little or no water (After Price, 1987).

and sands are considered permeable, whereas silts, clays or unfractured rock are of low permeability or impermeable. Permeability is an intrinsic property of the porous medium i.e. it does not depend on the nature of the fluid flowing through

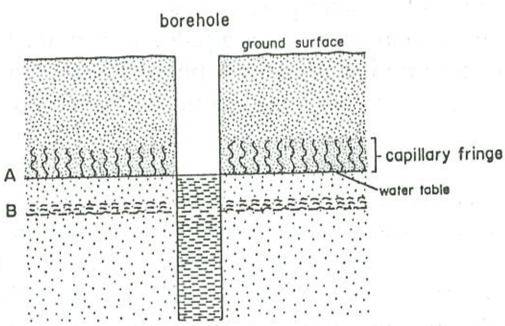

Figure 10.2 The capillary fringe. If the water table is lowered from A to B the capillary fringe will also move down. If the aquifer material at level B is coarser than at level A, then the capillary fringe will be thinner (After Price, 1987).

the medium. The units of permeability are cm^2 and values may range from 10^{-3} for a coarse gravel down to 10^{-15} for a fine glacial clay.

In areas where precipitation can percolate down from the surface to the water table, air and water are in contact and pressure at the water table is atmospheric. Such an aquifer is unconfined (Figure 10.1). New Zealand examples of unconfined aquifers are parts of the Aupori peninsula in Northland, the Pauanui spit on the eastern Coromandel coast, the Hamilton basin and much of the Wairarapa and Canterbury plains.

If however, impermeable material is close to the surface and overlies an aquifer full of water, air and water are not in contact, and the water pressure will usually be greater than atmospheric. Such an aquifer is said to be confined. Impermeable material may lie over or beneath the aquifer, forming confining layers. Water can only enter a confined aquifer by flowing laterally from an unconfined area. Examples of confined aquifers are the Kaawa shellbeds between Manukau harbour and Pukekohe, beneath the Rangitaiki Plain and much of the Heretaunga plain of Hawkes Bay.

The Ability of a Porous Medium to Store Water.

If a cubic metre of saturated material could be taken undisturbed from an aquifer, water would drain from it under the force of gravity. The volume of drained water, expressed as a percentage of the bulk volume (1 cubic metre) is called the specific yield of the material. Not all the water will drain from the material, some is held within the pores by capillary forces - this percentage is the specific retention. The sum of specific yield and specific retention is equal to the porosity.

If water enters an unconfined aquifer, the water table rises as the pores are filled. Porosity varies widely in different aquifer materials, up to around 45% in uniform- sized sands or gravels, so a considerable volume of water can be stored in an unconfined aquifer. For example, if the water table rises 1 metre in an unconfined aquifer with an area of 100 ha (10^6 sq metres) and a specific yield of 30%, then 300,000 cubic metres of water has gone into storage.

Conversely, if an aquifer is confined, it is full of water and has very little capacity to store more. Extra water can only enter a confined aquifer if the water pressure increases. This inflates the aquifer very slightly and compresses the water by a minute percentage. These are elastic (ie reversible) changes.

The combined effect of inflating the aquifer and compressing the water gives rise to a small but measurable increase of storativity (sometimes called storage coefficient). Storativity is defined as the amount of water taken into or released from storage in an aquifer per unit surface area of aquifer for each unit change of water pressure head. It has no units. Thus the definition applies to both unconfined and confined aquifers, but the numerical value of storativity in an unconfined aquifer may range up to 0.3 whereas in a confined aquifer it may be as small as 10^{-4} to 10^{-6}.

Transmission of Water through a Porous Medium.

A small volume of aquifer material may consist of mixed sizes of sand and gravel, it may be layered, or perhaps be a solid but fractured rock. At the scale of centimetres the aquifer material is discontinuous, the physical properties may be different in different directions and the material is said to be inhomogeneous and anisotropic. For a volume of aquifer material which is much larger than the largest particle, or layer, or with dimensions much greater than the largest spacing between cracks, the effect of the many discontinuities can be averaged and the material can be considered homogeneous, though it may still be anisotropic if it is layered. If an aquifer material is homogeneous and isotropic, it is much easier to develop theories to describe mathematically the movement of water through it.

The technical term which describes the ability of an aquifer to transmit water is transmissivity. The formal definition of transmissivity is the rate of flow of water through a unit width of aquifer under unit hydraulic gradient. The flow is assumed to occur through the full thickness of the aquifer, and the mathematical definition of transmissivity T is:

$$T = Kb \tag{10.1}$$

where K is the hydraulic conductivity (Chapter 8) and b is the thickness of the aquifer. The units of transmissivity are m^2/day or m^2/sec. Transmissivities of 30,000 m^2/day have been measured in the Heretaunga plains gravel aquifers of Hawkes Bay but this is exceptionally high. More typical values for rocks which can be considered to be aquifers are tens or hundreds of m^2/day.

Groundwater Movement

Water, whether above or below ground and free to move, always moves from a position of high energy to one of lower energy along the line of maximum energy gradient. In groundwater science this energy is referred to as the potentiometric or piezometric head.

Potentiometric Pressure, Flow Lines and Flow Rate.

Considering a point in water standing or with negligible velocity in a porous medium. The presence of the medium has no effect on the energy state in this case because there are no energy losses from friction between the porous medium and moving water.

However as fluid moves slowly through a porous medium, viscous friction causes a slow loss of energy, and the total energy line slopes. This slope is called the hydraulic gradient.

At any point in a confined or unconfined aquifer the water has a particular potentiometric head. If all these potentiometric points are joined they form a potentiometric surface, and it is possible to draw contours on this surface. The surface may at times be quite irregular (Figure 10.3) and so therefore will be the contours. A curved line drawn across the potentiometric contours and intersecting them all at right angles represents a line of maximum hydraulic gradient. Water entering the aquifer at the point of highest head on this line will flow along it, in the same way that a ball released on an irregular hillside will roll down the line of steepest slope. Many lines drawn in this way constitute a flow net and if this is done systematically useful general information on flow patterns can be derived (Figure 10.3).

The space between two flow lines is called a flow tube. Because each of these lines represents the path of all particles of water on it, no water enters or leaves a flow tube across these lines, though in an unconfined aquifer some may enter from the surface and some leave by springs or by being pumped out. A flow net is a useful tool in visualising the whole flow system and determining the relative transmissivity from one part of the tube to another.

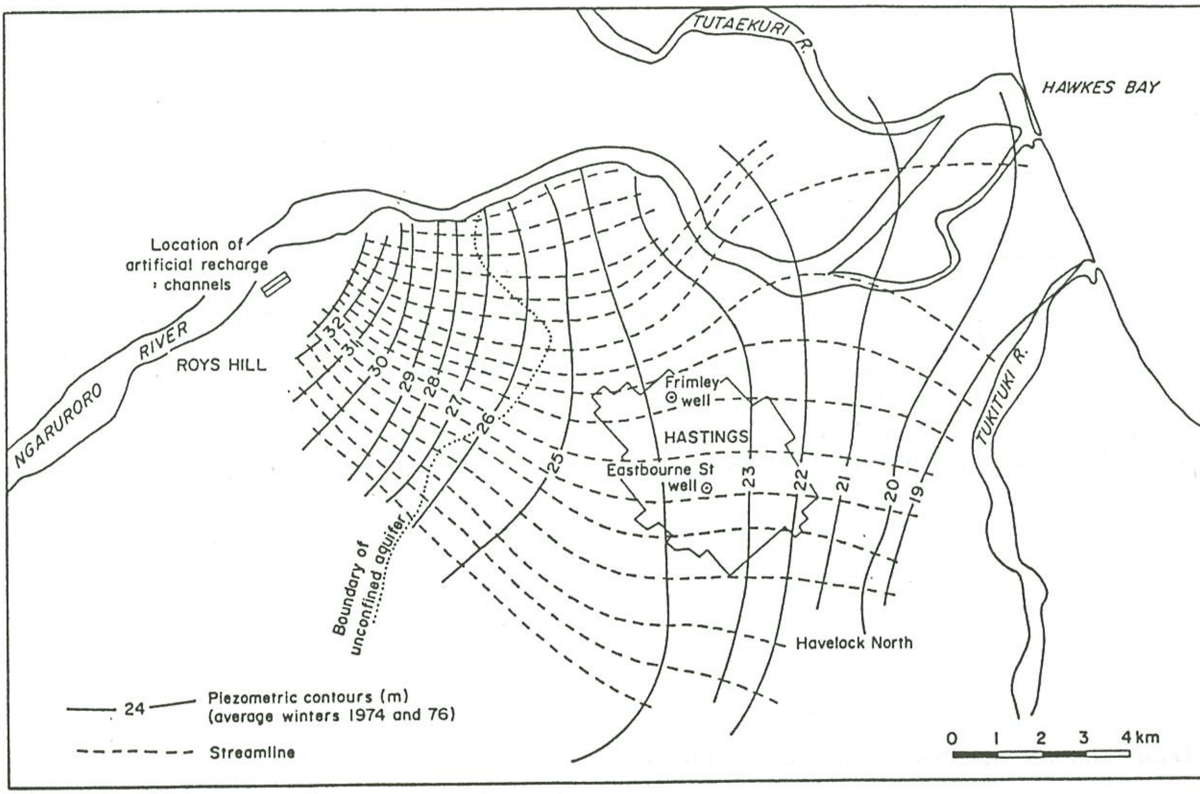

Figure 10.3 Flow net of a portion of the Heretaunga Plains also showing the location of the artificial recharge channels (Thorpe et al, 1982).

Darcy's Law.

Darcy's law is used to calculate the velocity and volumetric rate of flow through a saturated porous medium.

Henri Darcy was a civil engineer who was interested in supplying water to the city of Dijon, France. As a result of his experimental work he published the law which bears his name:

$$Q = AK.\frac{H_1 - H_2}{L} \qquad (10.2)$$

where Q is volumetric rate of flow, A the cross sectional area of the porous medium through which the flow is passing, K the hydraulic conductivity, H_1, H_2 are the potentiometric heads at the beginning and end of the flow path and L is the length of the flow path. K has units of m/sec (or cm/sec) and values range from 1m/sec for a coarse gravel down to 10^{-12} m/sec for a fine glacial clay. These values are determined both by the properties of the porous medium and of the water flowing through it. This is why it is called the hydraulic conductivity. Hydraulic conductivity is related to intrinsic permeability k by the equation:

$$K = k.\rho\frac{g}{\mu} \qquad (10.3)$$

where ρ and μ are the water properties of density and dynamic viscosity and g is the gravitational acceleration.

Since Darcy derived his law when experimenting with water filters, this equation can be illustrated by a simple filtration experiment (Figure 10.4).

In any channel or pipe with a cross section of area A carrying flowing water, the flow rate

$$Q = A.q \qquad (10.4)$$

where q is the average velocity of flow.

Comparing the equations 10.2 and 10.4 it can be seen that the flow velocity

$$q = K\frac{H_1 - H_2}{L} \qquad (10.5)$$

$\frac{H_1 - H_2}{L}$ is the hydraulic gradient often written as $\frac{dH}{dx}$ where x is the distance in the direction of flow and dH, dx can be read as small increments of H and x.

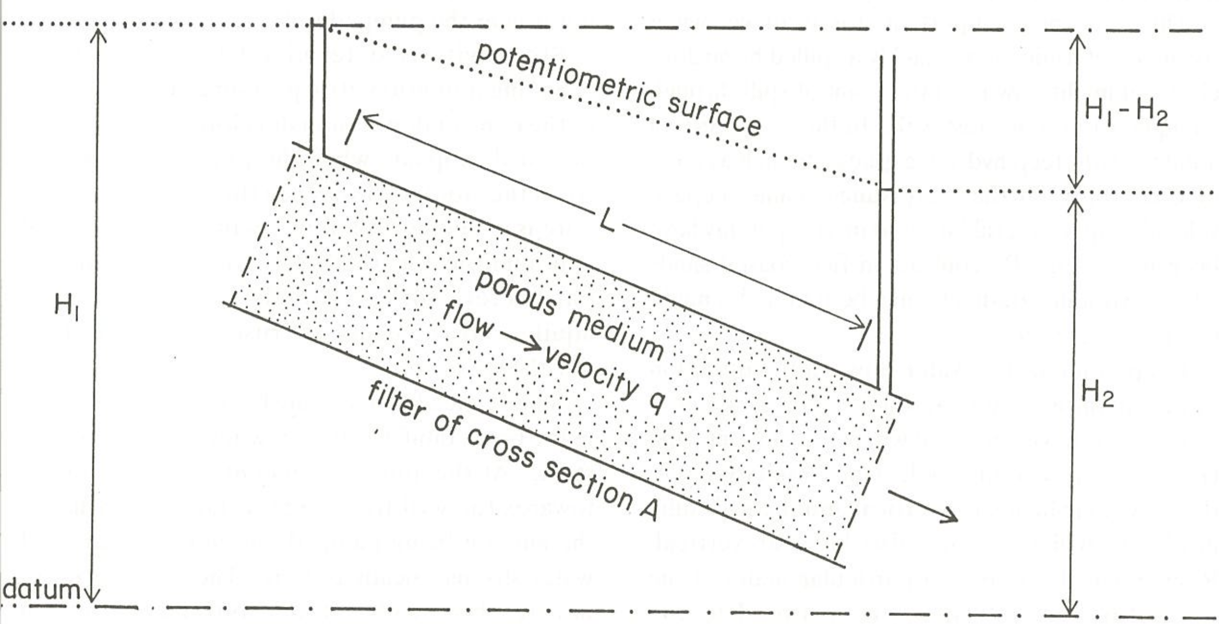

Figure 10.4 Schematic of a simple filter experiment demonstrating Darcy's Law.

Darcy's law is often written in the shortened mathematical form of

$$q = -K \frac{dH}{dx} \quad \textit{(See chapter 8)} \qquad (10.6)$$

where the minus sign just indicates that H decreases as x increases.

When Darcy performed his experiment he measured the cross section of his filter, ignoring the fact that much of the flow path was occupied by sand. The velocities he derived are thus based on this assumption and it is adequate for many engineering calculations. However because of the presence of the grains of the porous medium, the true total flow cross section is much smaller and therefore the average velocity of water between the grains must be higher (Equation 10.4). This is simply calculated as the seepage velocity Vs where

$$Vs = \frac{q}{\Theta} \qquad (10.7)$$

Θ being the porosity, a number less than 1, usually in the range .15 to .45.

This concept is important, for instance, when trying to determine how quickly a spilled hazardous chemical might flow from the point of spill through an aquifer to the nearest well. In the coarse gravel aquifers with steep hydraulic gradients such as parts of Canterbury on the Heretaunga plains seepage velocities up to several hundred metres per day have been measured. By contrast in fine coastal sands where hydraulic gradients may be flatter, Vs might be metres per year.

Keep in mind that water flows in the direction of maximum hydraulic gradient. The gradient is usually near horizontal, but where water is entering (recharging) an aquifer or leaving (discharging) it, the flows may be nearer vertical and the hydraulic gradients will therefore also be near vertical. Remember also that in a particular material the water flow velocity q is proportional to the hydraulic gradient (Equation 10.5).

Flow to a Well.

Darcy's equation is still applied to many practical engineering problems and one of the most common of these is the extraction of water from a well. However, Darcy experimented with homogeneous isotropic sands whereas a natural aquifer material is almost always heterogeneous and non-isotropic i.e. it may be layered or fissured. Nevertheless for such calculations the ideal situation is assumed, together with other simplifying assumptions, and the methods work well enough for practical purposes.

If the hydraulic gradient is very flat the water flows slowly. The action of pumping lowers the water level within a well and increases the gradient towards it (Figure 10.5) so that water velocities increase radially inwards. The overall effect is that a cone of depression, centred on the well, forms in the potentiometric surface. The difference between the original pressure heads and the reduced heads in and around the well is called drawdown. This phenomenon can be used to calculate the properties of the aquifer, transmissivity and storativity, by measuring drawdown in observation bores near the pumped well.

Storativity is determined by studying the changing potentiometric pressure or water levels as the cone of depression develops during pumping, or disappears when the pumping stops, i.e. when the storage of water in the aquifer near the bore is changing. There are numerous analytical and numerical methods for determining aquifer parameters for an equally large number of aquifer environments. (Kruseman and de Ridder, 1991).

Some time after pumping begins the water level in the bore stabilizes, the drawdown becomes constant. At this time the amount of water flowing towards the well from large distances is equal to the amount being pumped out and the change of water storage locally is zero. The flow is said to have reached steady state and its transmissivity can be analysed fairly simply.

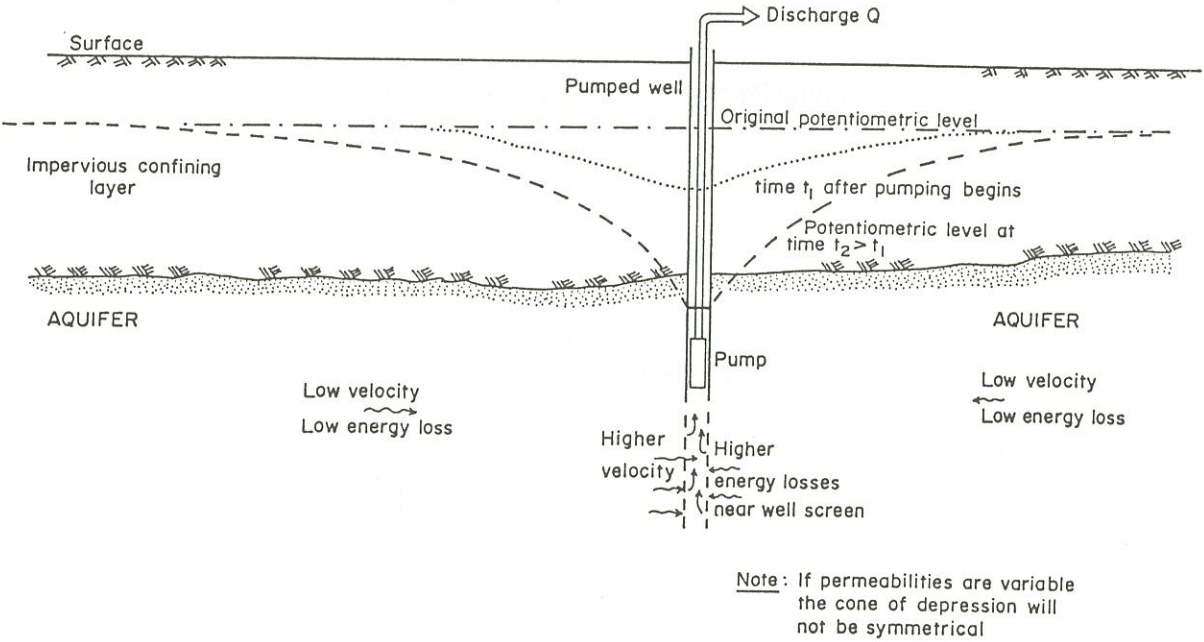

Figure 10.5 The creation of a cone of depression around a well during pumping.

Surface/Groundwater Interaction

Fresh water enters aquifers from rain falling directly on land overlying an unconfined aquifer, or by seepage from rivers and lakes. These are the processes of aquifer recharge.

Recharge from Rainfall.

The movement of water through the soil profile is covered in chapter 8 in detail but generally, water which penetrates below the root zone of surface vegetation (which may be several metres deep) will percolate downward until it reaches the water table.

The percentage of rainfall which recharges an aquifer depends on the type and thickness of the soil, the type of vegetation growing in it and the pattern of rainfall. For a given annual rainfall, more will recharge the aquifers if the falls are heavy and infrequent rather than in the form of light showers. Light showers tend to be soaked up by the soil and returned to the air by evapotranspiration, whereas heavy falls may overfill the moisture holding capacity of the soil and provide excess for recharge. Also, soils generally contain worm-holes, root holes or shrinkage cracks which in heavy rain provide flow paths through which water can bypass the upper zones of the soil and penetrate quickly beyond the level of most active evapo-transpiration.

Most recharge occurs between autumn and spring because evapotranspiration is low, soils are damper, and so their capacity to retain extra water is lower. Estimating the average annual recharge of an aquifer is very important for water managers because it is an upper limit of how much water can be taken from the aquifer on a sustainable basis. Figure 10.6 shows the pattern of natural recharge from rainfall over a 15 year period of the Winchmore Research Station near Ashburton. Over this period 31% of the rainfall passed through the soil profile to recharge the unconfined aquifer.

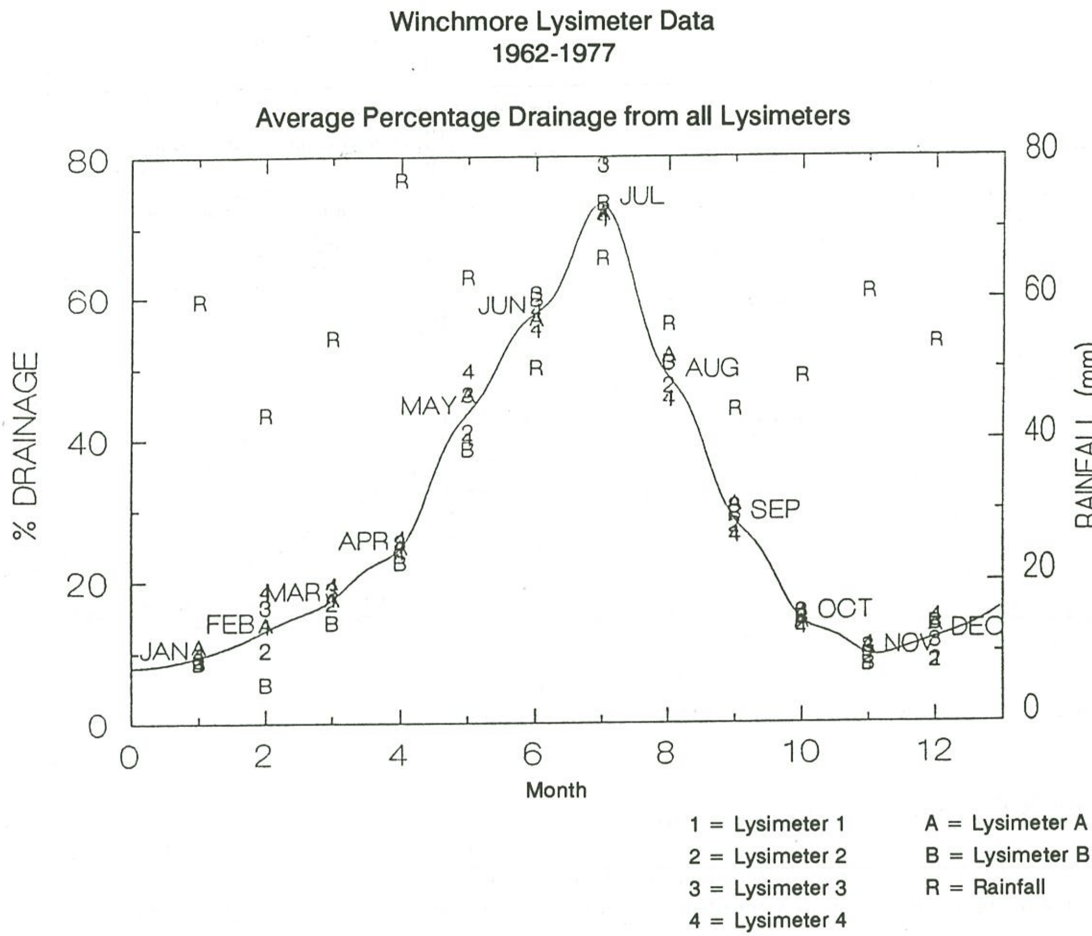

Figure 10.6 Distribution of recharge throughout the year at Winchmore Research Station, Mid-Canterbury (Courtesy of MAF Tech).

Ground Water Discharge.

In many instances ground water will discharge from an aquifer at a spring or perhaps into a swamp fed by dispersed groundwater seepage. On part of the Canterbury Plains (NCCB, 1986) recharge occurs at the higher points of the plains and discharge at the lower parts near and beneath Christchurch (Figure 10.7). Apart from the presence of springs and seeps, this can be deduced physically because when drilling a well the potentiometric pressure increases with depth (ie there is a hydraulic gradient vertically upwards). The opposite is usually found in recharge areas.

Artesian Flows.

Confined aquifers may be called aquifers when the pressure in the aquifer is high enough that the potentiometric level is above ground. Thus if a well is drilled into such an aquifer the water will rise to the surface and flow freely without being pumped (Figure 10.1). The coastal areas of the Heretaunga plains and Waimea plain near Nelson are like this as is also the area near and beneath Christchurch. In parts of the Kakanui Valley south of Oamaru the artesian head is 30 metres above ground level. The reason for this is that the recharge area for the deep confined aquifer is high in the hills to the northwest.

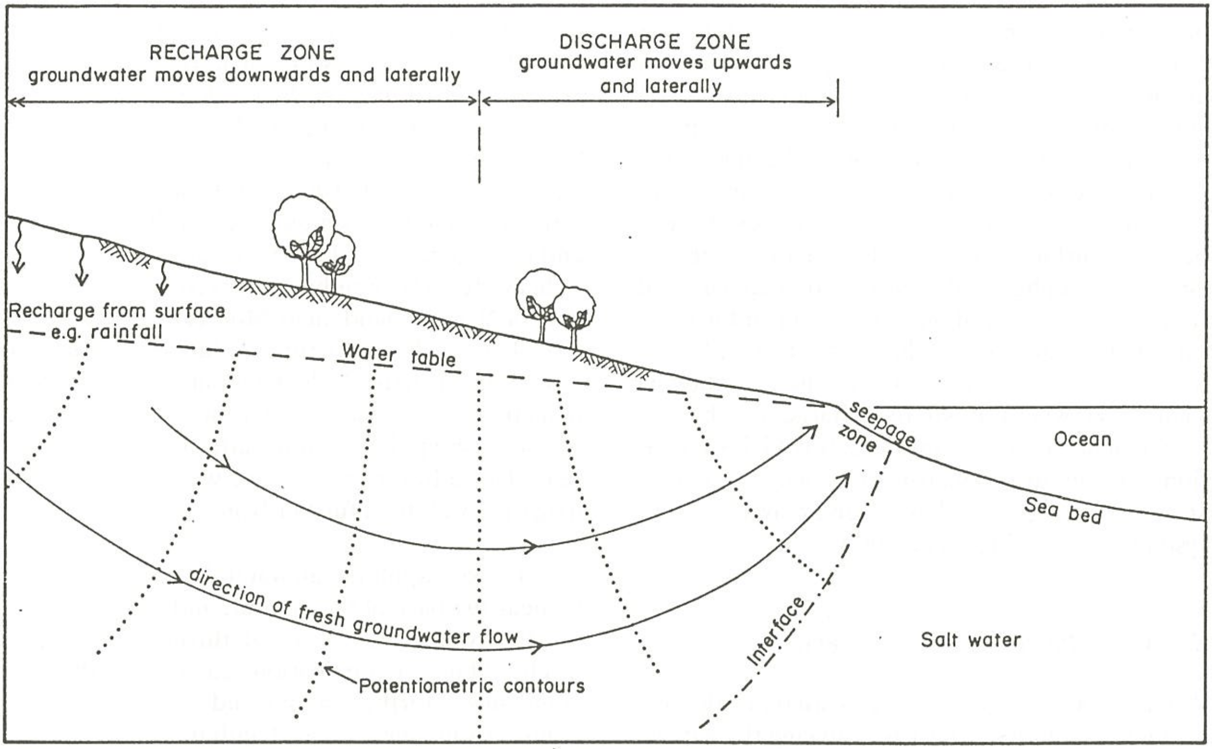

Figure 10.7 The classical concept of recharge to and discharge from an unconfined aquifer (NCCB, 1986).

Recharge From and Discharge to Rivers.

Recharge from or discharge to rivers can be determined by measuring flow changes along the river length. Conditions in the river must be low and steady because fluctuating flows occurring during even a small flood would mask the recharge/discharge effect. Measurements are made as near simultaneously as possible and the differences represent recharge or discharge to or from the aquifer over the river reach between flow gauging sites. Recharge and discharge may occur in different reaches of the same river as it flows across a plain, usually with recharge to the aquifers in the higher reaches and discharge further downstream. Examples of this are the rivers draining off the Mamaku plateau on to the Tokoroa plateau, the Wairoa-Waiiti Rivers near Nelson and the Ashburton river in mid Canterbury. Accurate measurement of river losses and gains is difficult because the changes are often small differences between much larger numbers which are masked by irreducible errors in the flow gaugings.

Measuring losses or gains from aquifers to lakes is more difficult again and usually can only be done qualitatively (i.e. is the lake gaining or losing), rather than determining actual flow rate.

Low flow in rivers is supplied by groundwater. Such flow arises from the steady seepage of water from the soil and fractured rocks into channels, mostly in the upper valley reaches of rivers. This water enters the ground during storms and is stored for varying periods up to years before being released.

Artificial Recharge

In areas where groundwater is heavily used, the natural recharge of the aquifer may be inadequate to meet the demand. This problem is addressed by introducing techniques of artificial recharge. These are often large basins or pits overlying an unconfined aquifer into which water is introduced at times of surplus such as floods or during winter, or alternatively low weirs across ephemeral river channels. Use is made of the fact that an unconfined aquifer can store very large volumes of water per unit surface area. Such structures are engineered to enhance the infiltration capacity and require regular maintenance to keep infiltration rates high. This concept has so far been adopted only twice in New Zealand, in the Heretaunga Plains of Hawkes Bay where recharge trenches are used (Figure 10.3) and in the Levels Plains Irrigation Scheme, near Timaru, where large diameter shallow recharge wells have been excavated. Both systems are working successfully.

Sea Water Intrusion Into Aquifers.

Where an aquifer is open to an estuary or the sea, groundwater flows seawards following the natural hydraulic gradient (ie the potentiometric head under the land is higher than mean sea level). If extraction of groundwater near the coast is too great, the head may be drawn down below mean sea level, so reversing the hydraulic gradient. Saline water may then flow inland with adverse effects on irrigated crops and those who drink it. Thus coastal aquifers must be carefully monitored and managed to avoid excessive drawdown.

Seawater intrusion has occurred in at least one part of New Zealand, near Motueka where irrigation of orchards was disrupted. In this case the salt water appears to be flushed out during winter when irrigation ceases and aquifer pressures recover, re-establishing the natural pattern of flow to the sea. The solution in this case was to drill a new irrigation well field further from the coast (Figure 10.8).

In thicker aquifers, salt water, being denser, may lie near the base of the aquifer, and during winter the freshwater may flow out through the upper levels. Thus, each irrigation season the saltwater could move further inland and that part of the aquifer would have to be abandoned.

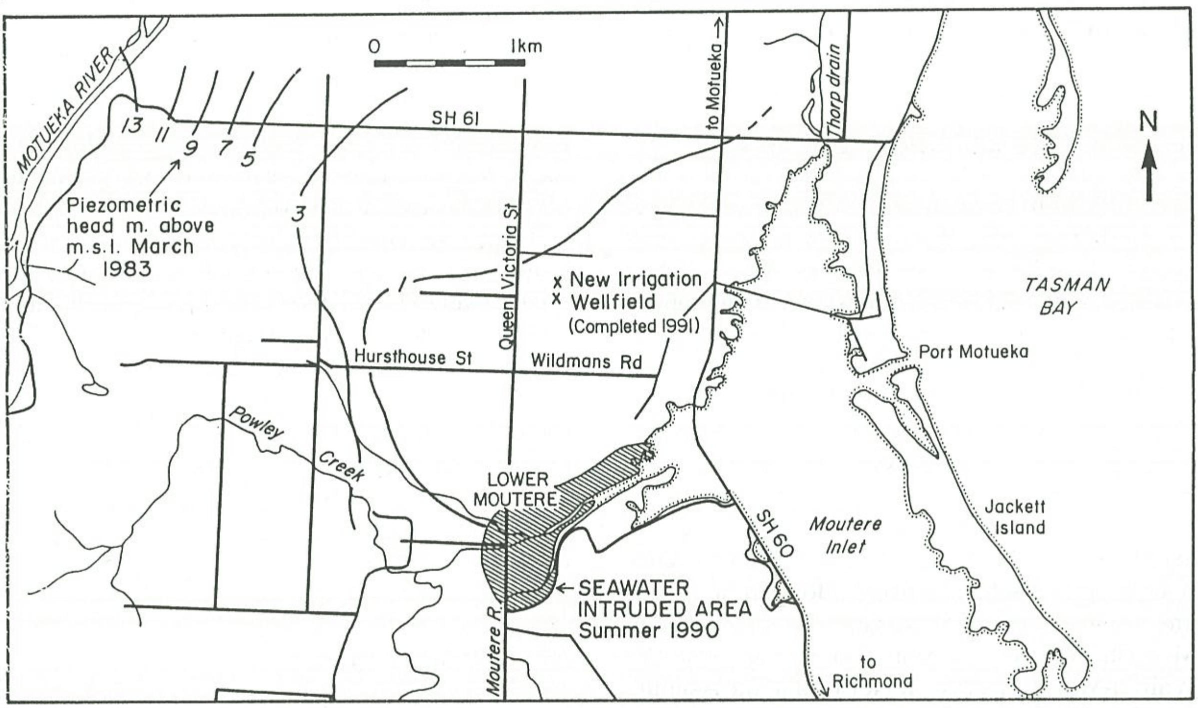

Figure 10.8 Sea water intrusion into a shallow aquifer at Lower Moutere, Nelson (Courtesy Fenemor and Thomas).

Three New Zealand Aquifers

Aquifers of New Zealand are generally quite small, the largest being the alluvial gravel aquifers of the Canterbury Plains. Nevertheless they are usually complex, partly because this is such a tectonically active country, and they include a great variety of types. Three will be briefly described here.

Aupori Peninsula.

The Aupori Peninsula stretches north from Kaitaia, towards Cape Reinga. There is a basement of volcanic rock at variable depth which is overlain by a Pleistocene sedimentary sequence. At the base of this sequence there is in places a shell bed and above this a variable sequence of fine sands containing silty sands and thin peat layers (GCNZ 1987). Sands of various ages occur at the surface, the younger sands forming recent dunes.

Horticultural activity increased in the Houhora area in the early 1980's creating a demand for irrigation. Somewhat earlier, pines were planted extensively on the west coast and these reduce ground water recharge because of their increased interception. The peninsula is narrow so there also is the potential that over-extraction of ground water might lead to intrusion of sea water beneath the land.

These considerations led to a study of the ground water resource before development proceeded too far.

The core component of the investigation was a transect of five investigation bores drilled across the peninsula in a SSW direction from Houhora. Piezometers were installed in these bores, and readings of potentiometric pressures from these allowed the flow system to be interpreted (Figure 10.9).

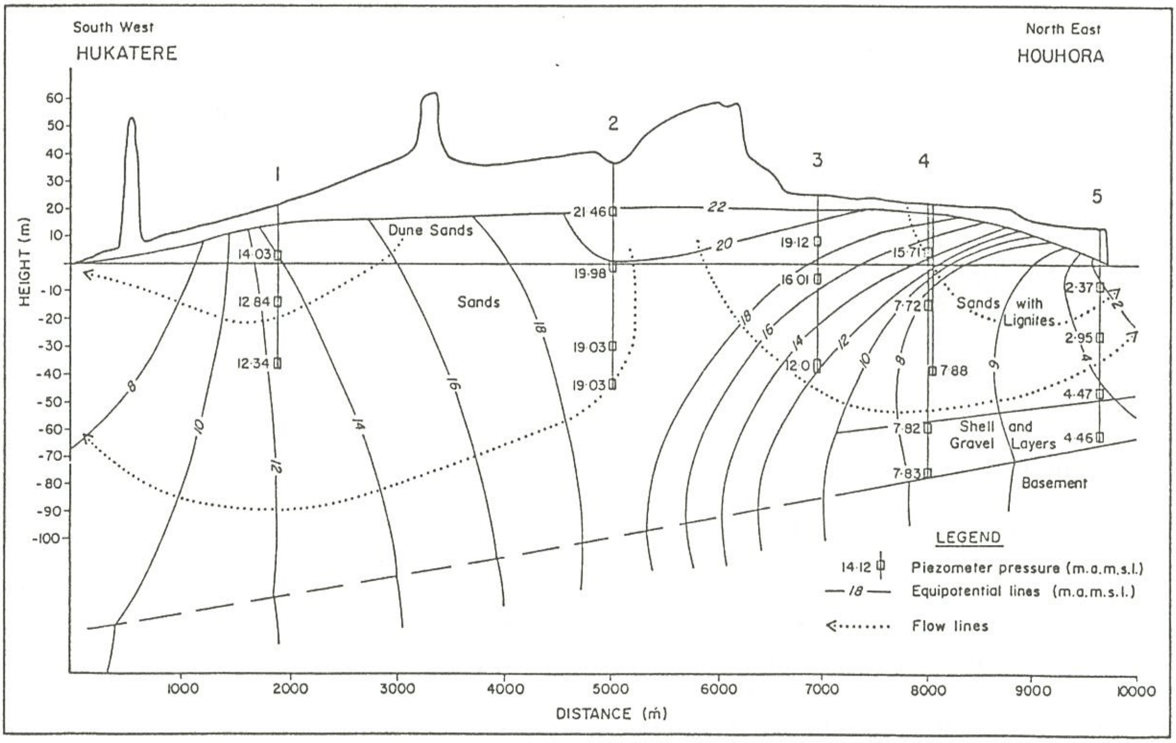

Figure 10.9 The hydrogeological cross section of the Aupori Peninsula near Houhora, Northland (GCNZ, 1987).

The only possible source of ground water recharge is infiltrating rainfall and the piezometers confirmed a mounded water table beneath the higher terrain, sloping to near sea level at the coast. The piezometers also showed pressures decreasing with depth in the central area of the peninsula (ie a recharge zone) and pressures increasing with depth near Houhora (ie upward flow in a discharge zone).

Hydraulic conductivities of the sands were measured and simple applications of Darcy's Law allowed a first estimate of ground water flow rates to be made which was confirmed by a simple recharge estimate using the method of Coulter (1973). There were significant differences between estimates of the steady state flows towards the coasts, but the computations suggested that a useful resource existed. The size of the ground water mound in the central area also showed that a large volume of water was stored in the unconfined aquifer which could be drawn on to supply seasonal irrigation requirements, provided that potentiometric heads were maintained above mean sea level in the coastal areas.

The Canterbury Plains.

The Canterbury Plains in total cover about 8000 km^2 built up from coalescing fans of gravel carried by rivers originating in the Alps. The gravels constituting the Plains aquifers are indistinctly layered and range up to 500m thick. Groundwater is not available everywhere as the gravels are mixed with variable amounts of sand, silt and clay. The layering is believed to be of material deposited during successive glacial and interglacial periods with the surface layer being post glacial. During glaciations, the increase in volume of material eroded from the Alps was such that the rivers could not transport it all and the channel slopes increased. During non-glacial periods the gravel supply decreased and the rivers were able to remove it from the upper plains and re-deposit it on the lower plains. During this process fine materials were selectively removed so that the transmissivity of the gravels increases with distance from the mountains. The shallower gravels are generally more transmissive than the deeper material.

With time the plains have grown larger, and the gravel formations thicker, covering a larger area and lying out across former sandy sea floor. In addition, sea level has been much lower in glacial times and gravel has been transported further seaward than the present coast. When sea level rose in interglacial periods to levels similar to the present day, marine silts and clays were deposited on top of the gravels. The final result near the coast is a sequence of gravel aquifers separated by fine grained confined layers so that Christchurch sits on at least four aquifers which are flowing artesian in the eastern and southern parts of the city and from which an excellent water supply is drawn (Figure 10.10).

The aquifers are usually hydraulically connected with the rivers. Major rivers such as the Rakaia and Waimakariri are incised across the upper plains and do not lose water to (ie recharge) the aquifers until their lower reaches (Figure 10.11). The ground water which Christchurch uses enters the aquifers from the Waimakariri River beginning about 30 km from the coast. Over a reach of some 25 km it is estimated (Wilson, 1973) that 10-12 m3/sec on average leaves the river. Some becomes under flow i.e. flowing through the gravel bed parallel to the river but some flows through the south bank of the river, recharging the unconfined aquifers. Conversely, smaller rivers such as the Selwyn lose water to the aquifers in the upper reaches and regain it in the lower reaches.

In areas of the plains away from rivers the recharge to ground water is from excess rain infiltrating through the soil. This infiltrating water carries with it dissolved chemicals such as nitrate or sulphate which can be detected and help to define the areas recharged by rain from those recharged by rivers. Present estimates are that this recharge is on average about 30-40% of average annual rainfall (ie 250 - 325 mm per year).

Although Christchurch (population about 300,000) uses a large amount of ground water, the greatest use in Canterbury is for irrigation. The often droughty soils of the plains and the tendency to hot dry summers create a great demand for water, but as much as 60% of the irrigation water

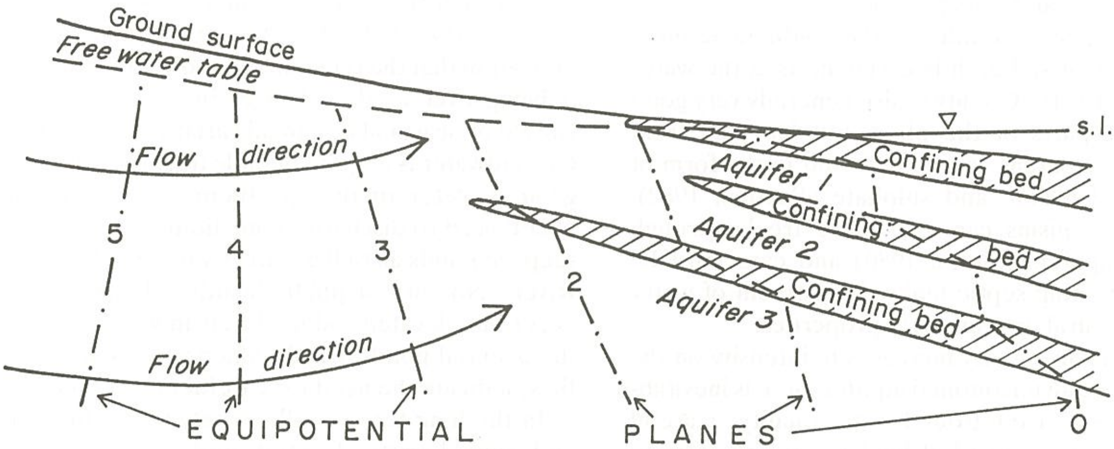

Figure 10.10 A schematic of the aquifer system beneath Christchurch (Wilson, 1976).

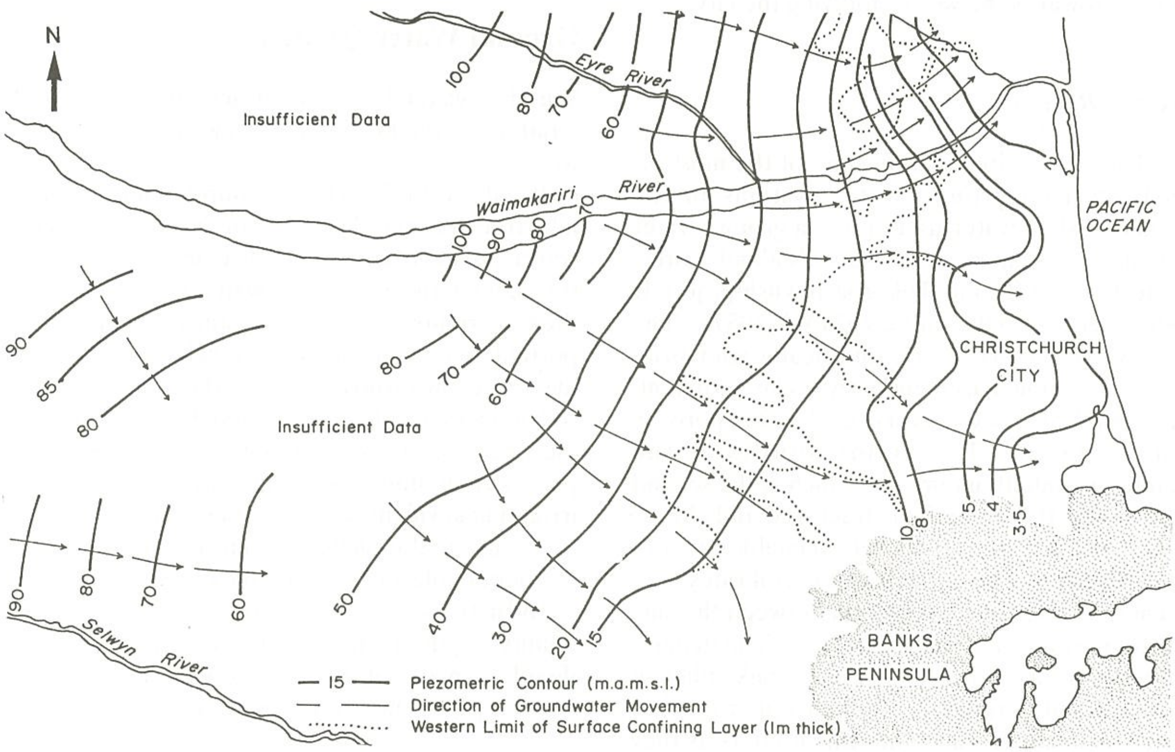

Figure 10.11 Geohydrology of the Canterbury Plains near Christchurch. Major recharge of the unconfined aquifer occurs through the south bank of the Waimakariri River from about the 60m contour upstream (NCCB, 1986).

applied is recycled through the soil to ground water (Scott and Thorpe 1986).

Ground water quality in the confined aquifers beneath Christchurch is excellent as is the water near the rivers. Quality is also generally very good in other parts of the plain, but leaching from agricultural areas can be detected in the form of increased nitrate and sulphate (Burden, 1982). Micro-organisms can also move freely through gravel aquifers (Sinton 1980) and care must be taken in siting septic tanks downstream of water wells on rural or semi-rural properties.

As human activity increases in intensity on the land above an unconfined aquifer there is inevitably some effect on ground water quality. Careful planning and control of land use is necessary to minimise these changes. Areas on the northern and western margins of Christchurch are especially important as ground water from these areas flows on towards the wells supplying the city.

Tokoroa Region.

The Tokoroa region contains one of the most intensively used ground water systems in the Waikato. Most water users rely on ground water and the two major users, New Zealand Forest Products (NZFP) and Tokoroa township, jointly abstract about 35,000 m^3/day (WVA 1987). The geology of the area is a series of volcanic ignimbrite sheets overlying basement greywacke sediment. The ignimbrites are of varying density, porosity and hardness. The major aquifer is the Waiotapu/Kinleith ignimbrite which is the second layer down. It is extensively fractured and closely jointed, and it is these joints which enable water to move readily and be extracted at useful rates.

There is extensive interaction between the surface and ground waters in the area. The general pattern of stream flow is from the Mamaku plateau westward and northward. In their upper reaches these streams lose water into the aquifers as they descend from the Mamaku plateau but much spring water re-enters the streams as they further descend from the Tokoroa plateau towards the Waikato river (Figure 10.12). Recharge also occurs from rainfall but may take several months to have an effect on ground water levels. Estimates are that half the average annual rainfall becomes recharge (WVA 1987). At present there is no indication that the regional ground water resource is being over used. Changes in water level are caused by seasonal or annual variations in rainfall. Groundwater is of good potable quality. Potential ground water quality problems exist because NZFP used to discharge some liquid effluent into seepage ponds as well as directly into the Waikato River. No current public health risk appears to occur but elevated sodium levels in wells north of the disposal ponds, the direction of ground water flow, indicate the need for careful monitoring.

In the long term, conflicts between industrial- and animal-waste disposal, and potable use of ground water, may be the most significant resource-use conflict in the region.

Ground Water Quality

Whether water be from surface sources or from aquifers, its quality determines what it can be used for.

If water is for human consumption it is vital that it be free of harmful bacteria or viruses since these water-borne organisms cause many epidemics throughout the world. If water is for irrigation then the removal of micro-organisms is not so important, but chemical quality is relevant because it may affect the growth of plants, the structure of the soil or operation of an irrigation system. For instance the presence of boron at concentrations of 1 or 2 parts per million makes the water unsuitable for irrigating kiwifruit, and too much sodium in irrigation water deflocculates certain soils, making them less permeable and less suitable for growing plants.

Industries may also have particular water quality requirements. If water contains too much dissolved material this may deposit as a scale on interior pipes and reduce boiler efficiency.

Natural ground water chemistry.

Ground water quality may be determined from natural causes or may be affected by human ac-

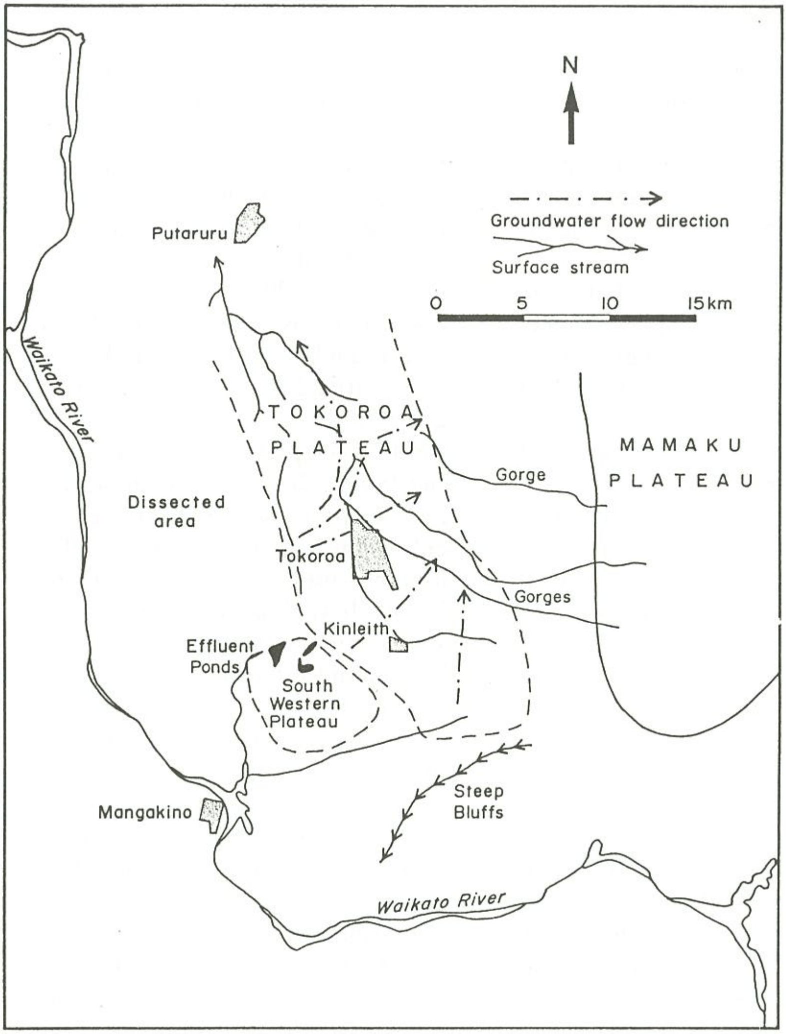

Figure 10.12 The surface and groundwater flow paths of the Tokoroa area (WVA, 1987).

tivity. If the human effects are troublesome the water is said to be polluted or contaminated and this could be from leaking petrol storage tanks, poorly located septic tanks, spills of toxic chemicals from tanks or tankers, or just from intensive horticultural activity.

When rain strikes the ground it is already carrying a very small chemical load, typically 10 to 20 mg/*l* (parts per million). In coastal areas this would be higher because the winds from the sea carry spume over land and the rain from that direction has more sodium chloride.

As it percolates through soil, especially the biologically highly active root zone, the rain water reacts chemically with the soil particles, organic matter and soil gases. Absorption of carbon dioxide from the soil gas (produced by root and microbial respiration) is particularly important because this forms weak carbonic acid in the water which is then able to dissolve any carbonates present.

The chemical content of water below the root zone will also be different because of the process of evapotranspiration (Chapter 8). Since a percentage of incoming rainfall evaporates back into the atmosphere, the concentration of chemicals in the remaining water increases. Thus, if half the water is evapo-transpired, the dissolved chemical content will double. This is one reason why dissolved minerals in ground water are so high in areas such as Australia, where evapotranspiration in many areas is much higher as a percentage of rainfall than in New Zealand.

The chemical nature of ground water below the root zone continues to change, but at a very much slower rate. The water and rock through which it moves reacts chemically very slowly, though in limestone areas the rate of reaction is somewhat quicker. Because ground water is in contact with rock for long periods, ranging from years to millennia, the changes in water quality may be quite marked even though the rate is slow.

The general sequence of changes is broadly understood. Initially the water contains bicarbonate and calcium plus oxygen as major constituents, plus nitrate and sulphate. Nitrate naturally derives from nitrogen fixing by legumes and sulphate from oxidation of sulphides in the soil, but both these may be supplemented by agricultural activity on the land including leaching of applied fertilisers.

As water moves away from the recharge zone oxygen is slowly depleted. If the water flows beneath a confining layer, conditions eventually change from chemically oxidising to reducing, and calcium and any magnesium ions in solution are replaced by sodium. The predominant bicarbonate ion slowly changes to predominant sulphate in the water and then finally to chloride. This whole process may take centuries and the water will only complete this natural process of transformation where the aquifer conditions are suitable.

Groundwater Pollution.

In areas of ground water recharge, aquifers are vulnerable to some extent to contamination by undesirable substances - the water is at risk of being polluted. The risk is greatest in areas of most intense human activity, where the water table is at shallow depth and the overlying material is coarse. Thus river flats and terraces where towns and cities are often built may be areas of particular concern i.e. Hastings, Lower Hutt, Christchurch.

The unsaturated material above the water table protects ground water from pollution to some degree. Many pollutants never pass below the soil horizons because they are adsorbed at that level and degraded or transformed by soil microbial activity or other biological processes. Some degrade naturally.

A few persistent ions, notably nitrate nitrogen are very mobile and are found widely distributed in unconfined aquifers, generally as a result of agricultural activity. Conversely, where an aquifer does become polluted this may go unnoticed for a long time. Because ground water movement is slow compared with surface water flow, aquifers seldom flush themselves clean of pollutants, although some coarse alluvial gravels in New Zealand have done this with small spills of hazardous chemicals. Therefore it is vitally important that groundwater pollution is prevented rather than cured.

The key to protecting groundwater quality is wise use of the land above the aquifer, prohibiting or managing those activities which produce polluting wastes or store and use toxic chemicals in large quantities. Examples are timber treatment plants, service stations with buried storage tanks, meat processing plants, piggeries, factory poultry farms or fertilizer works. Incidents, some accidental, some resulting from normal operations, resulting in groundwater contamination have occurred in recent years near Masterton and Rotorua (timber treatment), Hamilton and Southland (dairy factories) in the Hutt Valley (leaking petroleum storage tanks) and Ashburton (freezing works).

Where a polluting material originates from a point source such as a piggery it is usually easier to recognise and to manage. Nitrate, however, is an all pervading pollutant arising from many different forms of agricultural activity. Even pastoral farming raises nitrate levels significantly and where nitrogenous fertilizers are applied a proportion will always leach into an unconfined aquifer. Generally, the more intensive the agricultural activity the higher the nitrate levels will become. As the nitrate originates over large areas it is exceedingly difficult to control, because there may be many different properties engaged in differing farm or garden activities. Fortunately nitrate is not highly toxic, but it is widespread in New Zealand ground waters. The limit for nitrate nitrogen in water set by the World Health Organisation is $10 gm/m^3$. Intensive pastoral farming in parts of Canterbury has raised nitrate nitrogen levels to $5 gm/m^3$ or more and intensive dairying activity in the Hamilton basin has resulted in a mean value of $11.7\ gm/m^3$ (Petch, pers. comm.).

Waste disposal is another human activity with the potential for seriously polluting ground water. Urban areas dispose of solid wastes in landfills which are designed with capping layers and bottom liners to minimise entry of water and control movement of water which does enter. Water will enter even well designed and managed landfills to some extent and produce highly polluting leachate. Thus it is most important that sites for new landfills be carefully selected to aid water control and minimise risk from escaping leachate.

Similarly it is becoming popular to dispose of liquid wastes from rural industries, such as dairy factories or from sewage treatment plants, by spraying on to land. This has many positive attributes as a waste disposal procedure but it must always be remembered that there may be an unconfined aquifer below the surface which could be polluted. Examples of sewage effluent disposal by spray irrigation of pine trees are the holiday settlements of Whangamata and Whiritoa on the Coromandel Peninsula, Rotorua city and Levin in the Horowhenua district. Freezing works near Christchurch and dairy factories in the Waikato and Southland also irrigate effluent onto pasture but with noticeable increases in nitrate nitrogen concentrations.

Clues from groundwater chemistry.

The chemistry of ground water is often used as an indicator of its origins and/or as a tracer of movement through the aquifer. In large systems with slow moving water, where the natural sequence of changes of chemistry has time to occur, the natural quality can be interpreted in this way. Nitrate levels above one or two parts per million indicate that leaching from agricultural activity has had an effect.

Natural isotopes of oxygen and hydrogen are most useful in identifying the sources of groundwater in many parts of New Zealand. These isotopes have slightly different evaporation and condensation properties from their normal elements, so that they tend to separate when undergoing such transformations. The relative amounts of the isotopes are affected by whether condensation occurs, for instance, at altitude in the mountains or on a lowland area. If an aquifer is recharged partly from rivers draining alpine areas and partly by rain falling on lowland plains, the proportion of each type of water in a sample from a well can be reliably estimated by measuring the isotope contents. This has proven a useful technique in Canterbury (Taylor et al 1989) in the Rotorua area (Taylor et al 1977). The unstable isotope of hydrogen, tritium, has also been a useful tracer and a means of estimation of water "age". Most environmental tritium in the past has been created by atmospheric testing of thermonuclear weapons which began in the 1950s and ended in 1964. Thus water which has very low natural tritium levels is older than 1950. Incoming tritiated rainwater has reduced since testing stopped and the natural radioactive decay of tritium has reduced levels to the point where the technique is becoming much less useful.

On occasion specific tracers may be used to determine flow paths over short distances. These are injected, usually at a point, and water is sampled in wells or at springs to see if the tracer appears. This may allow estimation of both flow directions and velocity, but the method may be very uncertain and time consuming since sampling must be repeated many times until either tracer appears or it is concluded that the water has flowed elsewhere!

Micro organisms in groundwater.

Despite the fact that there have been numerous instances worldwide of epidemics caused by drinking polluted ground water, the behaviour of micro-organisms in aquifers is still poorly understood.

There is a general presumption that when for instance, sewage effluent is applied to a land surface that any pathogenic organisms will be destroyed by the natural processes in the biologically active root zone but this is not well researched. A fine grained soil is certainly a good protection above an unconfined aquifer, but coarser soils or soils containing root or worm holes or shrinkage cracks may allow polluted water to bypass most of this protection.

Research on New Zealand gravel aquifers has shown that micro-organisms can survive for several weeks and move freely. Experiments in Canterbury using tracer bacteria have shown travel distances of more than 900 metres (Sinton, 1980). In semi-rural areas where people use ground water for domestic purposes and also rely on septic tanks for waste disposal this research has been useful in setting minimum spacings of wells and septic tanks.

The increasing popularity of sewage disposal by irrigation of treated effluent and sludge disposal onto land will require further research in these topics to assist designers and ensure that ground water quality remains high in the face of increasing urban pressure.

Use of Groundwater

One of the great advantages of ground water as a resource is that the aquifers often contain large volumes of water that moves slowly. They are thus a type of reservoir and provide a steady though still finite water supply. The ground water reservoir can be depleted during dry periods provided that it is managed so that the long term average drawoff is small enough to allow refilling during wetter spells. In some New Zealand aquifers, such as the Waimea Plains of Nelson which are pumped heavily for irrigation, there is a summer drawdown followed by winter recovery. In the Canterbury aquifers which discharge into the ocean such seasonal variations in water table level may be superimposed on longer period changes caused by sequences of years with rainfall wetter or dryer than average. Sustainable management of a ground water resource entails extracting water so that the long term average use is less than the long term average recharge.

In more arid climates the average recharge may be very small and the value of the water so great that sustainable management is not attempted. The ground water resource is "mined".

Irrigation.

Worldwide, irrigated agriculture consumes by far the greatest volume of ground water compared to other uses and this is true in New Zealand also, although hard information is lacking. All taking of water for uses other than domestic, stock watering and fire fighting is controlled by the issuing of a water permit.

However the rate and amount specified in each consent is an upper limit and the actual use is seasonal and weather dependent. An estimate made for an irrigated area of 18000 ha between the Rakaia and Ashburton Rivers for the very dry 1981/82 season is 35.8 million m^3 which equates to 139 mm of water applied (Price, 1987).

Urban Groundwater Use.

Ground water supplies, either wholly or in part, the needs of more than one third of the people of New Zealand. There are no data on rural use of ground water for domestic purposes but this would be minute compared to the urban areas.

Major cities and towns which are wholly dependent on groundwater are Napier, Hastings, Lower Hutt, Blenheim and Christchurch. The per capita consumption of water (including industrial use) in Christchurch averages 450 l/day (Crosby pers comm). On a hot nor-westerly summer day, garden watering may lift this consumption as high as 1500 l/day. Other cities such as Papatoetoe and Palmerston North are partially dependent. Towns like Greymouth and Timaru take water from large horizontal pipes set in gravels close to rivers so that it is prefiltered before entering the reticulation system, while still others such as Whangarei and Rotorua are partially dependent on spring water which has only just ceased being groundwater!

Table 10.1 Populations of territorial local authorities, townships and urban areas which are dependent wholly or partly on ground water (1989 population estimates 000's).

Name	Wholly Groundwater	Surface/ Groundwater
Ashburton	-	24.6
Balclutha	4.2	-
Blenheim	23.1	-
Cambridge	6.0	-
Carterton	-	6.0
Christchurch	257.6	-
Cromwell	3.5	-
Foxton	4.2	-
Gisborne	-	7.6
Gore	10.0	-
Greymouth	-	-
Hastings	55.6	-
Hope/Brightwater	2.5	-
Howick	-	20.0
Kaikohe	-	8.0
Lower Hutt	95.3	-
Lyttelton	3.2	-
Massey	6.0	-
Morrinsville	-	5
Mosgiel	9.1	-
Motueka	5.1	-
Napier	52.7	-
Nelson	-	45.4
Onehunga	16.6	-
Otaki	-	4.4
Palmerston North	-	66.8
Papatoetoe	-	30.0
Pukekohe	9.4	-
Putaruru	4.5	-
Rangiora	6.7	-
Rotorua	-	53.6
Selwyn	41.0	-
Te Anau	3.0	-
Temuka	3.9	-
Tokoroa	17.6	-
Wanganui	38.0	-
Wellington	-	150.0
Waipukurau	3.9	-
Waimakariri	-	44.0
Whangamata	2.5	-
Whangarei	20.0	-
Winton	2.1	-
Smaller Towns	23.6	-
Totals	727000.0	497000.0
% of Population of NZ	22%	15%

Industries.

Many industries in the cities and towns listed above are reliant on ground water but in addition major industries are based in rural areas to be close to sources of raw material or transport routes. The aluminium smelter in Southland has a resource consent to take 4500 m3/day of groundwater from the gravels below the Tiwai peninsula. A brewery and major food processing factories draw on the ground water below the Heretaunga Plains. New Zealand Steel uses about 1700 m3/day of water from the Kaawa shellbed formation at Glenbrook on the Manukau lowlands and major dairy factories in many parts of New Zealand also pump similar volumes.

Disposal of waste water into aquifers.

In semi-arid areas water has much higher value than is accorded in well-watered New Zealand. Water is used more efficiently and in some instances is re-used. A practice which is becoming accepted in many parts of the world is irrigation of treated sewage effluent on to land. In New Zealand this is regarded merely as a means of disposal with minimal risk to the environment and a possible side benefit of improving tree crops or stabilizing sand dunes (eg Levin). However in Israel and Arizona, to name two places, the process is regarded as one of renovation of water quality so that it can at least be used for further irrigation of crops and with further research and experience it may be considered acceptable for human consumption. It is in fact a form of artificial recharge and potential effects on groundwater quality must always be kept in mind.

Acknowledgements

Advice, information and editorial assistance from the following are gratefully acknowledged. D.L. Murray, P.A. Mosley, E.R. McSaveney and R.A. Petch. MAF Tech Winchmore Research Station provided the data from which Figure 10.6 was derived and A.D. Fenemor and J.T. Thomas provided Figure 10.8. Figures 10.1 and 10.2 are used with the kind permission of M. Price.

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11
Karst Hydrology

P W Williams

The word *karst* comes from a limestone region at the head of the Adriatic Sea between northern Italy and Slovenia (Yugoslavia), which in Roman times was called the Carsus. The landscape is unusual, even bizarre, with large rivers disappearing underground, caves, enclosed depressions, periodic lakes, subterranean rivers (hence the River Styx of classical legend), and large springs. These features became known as "karst phenomena", and this term is now applied to similar features and landscapes all over the world. New Zealand has a number of regions with karst terrain that are described in Williams (1992). Leisure activities and adventure sports give New Zealanders first-hand contact with the karst underworld through visits to caves, "black water rafting", and abseiling into features such as the Lost World and underground Mangapu River near Waitomo.

About 25% of the world's population depends on water from karst systems for their domestic, agricultural and industrial needs (Ford & Williams 1989). Some small Pacific countries, like Niue, obtain all their water from carbonate rocks. New Zealand is much less dependent on karst waters, although the carbonate rocks in which they are found are quite widespread (Figure 11.1) and contain potentially important water resources for the future.

Karstification

Karst is found in limestone and marble (metamorphosed limestone), because these rocks, being composed of calcium carbonate, are highly soluble in rainwater. Other soluble rocks such as gypsum also develop karst, but do not occur to a significant extent in New Zealand, so will not be considered further here. Rainwater contains dissolved carbon dioxide (CO_2),and is therefore a weak acid, carbonic acid (H_2CO_3). It becomes even more acidic as it percolates through the soil, where the proportion of CO_2 in the air in soil pores can be a hundred times greater than in the open atmosphere. This makes the percolating water very corrosive towards limestone. So as it seeps into fissures such as faults, bedding planes and joints in the rock, it gradually enlarges them by solution. In this process, the relatively insoluble calcium carbonate ($CaCO_3$) rock is transformed into the much more soluble calcium bicarbonate:

$CaCO_3$	+	H_2O	+	CO_2	=	$Ca\,(HCO_3)_2$
limestone	+	water	+	carbon dioxide		calcium bicarbonate

Over time, narrow fissures can develop into very large caves and the surface topography can become pocked with sinkholes, a process termed "karstification". Karst development occurs principally where carbonate rocks are hard and relatively free from impurities (less than about 20% clay); comparatively soft, marly limestones do not develop karst. Consequently, karst is not found in the argillaceous limestones near Silverdale just north of Auckland.

Karstification is a fundamental process in developing the water-bearing capacity of most car-

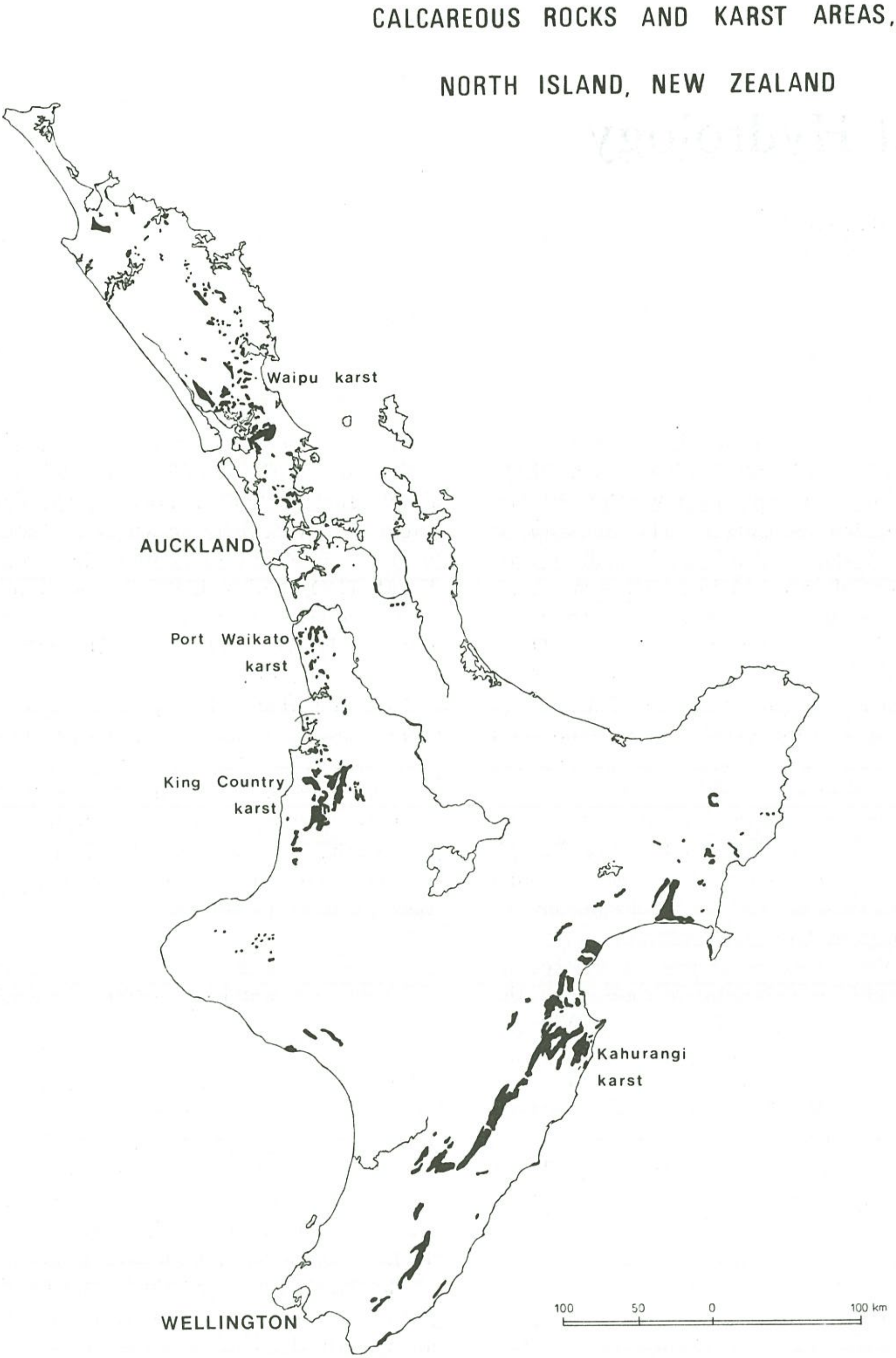

Figure 11.1a Distribution of carbonate rocks in the North Island (Source: New Zealand Geological Survey) (From Williams 1992)

CALCAREOUS ROCKS AND KARST AREAS,

SOUTH ISLAND, NEW ZEALAND

Figure 11.1b Distribution of carbonate rocks in the South Island (Source: New Zealand Geological Survey) (From Williams 1992)

bonate rocks. Interstices between the accumulating carbonate material as it is deposited form the primary pore spaces of the consolidated rock. Temporary halts in deposition produce bedding planes. Bending during tectonic movement produces cracks known as joints, and stronger movements cause major dislocations or faults. These fissures provide potential pathways for the movement of underground water, and can be thought of as the "primary porosity" of the rock. Karstification transforms this primary porosity, by chemical enlargement of the fissures, into a very much greater "secondary porosity". Circulation of water through the rock continually flushes the limestone with a dilute acid, and so the rock becomes progressively more and more riddled with interconnected voids of increasing size and complexity. The frequency of fissures that are sufficiently enlarged to permit rapid water flow through them increases with time, and some secondary voids may become extremely large. The largest natural underground void known is in the Mulu karst of Sarawak in Malaysia. It is a cave chamber with a volume of 20 million cubic metres that could accommodate about 15 football fields on its floor space! The largest chamber in New Zealand, in Hollow Hill Cave at Waitomo, has a volume great enough to contain an ocean liner. Such huge spaces are not produced by chemical solution alone, but by a combination of solutional enlargement by underground rivers, followed by repeated roof collapse and removal of the rock debris by river flow.

Karstification may not be uniform throughout a large body of rock, such as a limestone mountain. It tends to decrease exponentially with depth below the surface soil layer, which is the main source of CO_2 and hence carbonic acid. Percolating water is most corrosive to carbonate rock near the soil and becomes increasingly more saturated with bicarbonate as it percolates downwards. However, because of the unevenness of jointing and faulting, in some places water can penetrate more easily into the rock than in others,and hence move underground more rapidly and reach greater depths before becoming chemically saturated.

The uppermost, heavily corroded layer of rock beneath the soil is known as the subcutaneous zone or epikarst. Fissures are widest near the soil, but gradually taper or close with depth. Consequently, infiltrating rainwater flows into this zone more easily than it can flow out. As a result there is substantial water storage in the epikarst, especially after rain (Williams 1983).

Because New Zealand is a relatively wet country there is plenty of solvent, i.e. water, to dissolve karst rocks. In part of the Waitomo district with an average annual rainfall of 2350mm, the limestone is chemically eroded at a rate of about $69m^3/km^2$ each year (Gunn 1981). On Takaka Hill in Nelson, with an average rainfall of 2158mm, the marble is dissolved at a rate of close to $100m^3/km^2$ per year (Williams & Dowling 1979). At this rate, a kilometre long cave passage measuring $1m^2$ in cross-sectional area would take only 10-15 years to form, if all solution within $1km^2$ of outcrop were focused in one place. Of course this is not the case, but the simple calculation emphasizes that karstification is rapid in a wet country. In the few million years that New Zealand limestones and marble have been subjected to karstification, there has been ample time to develop the "plumbing" - the cave systems - of large mountains. For example, the uppermost cave passages in Nettlebed Cave beneath Mt. Arthur are more than 350m above the present water-table. Some of these passages, which were formed below the water-table but are now high and dry, are known from uranium/thorium and paleomagnetism dating to be more than 700 000 years old (Williams 1987). Hence, the many kilometres of intricate passages within the underlying 350m of marble have been formed in less than about 700 000 years.

Characteristics of Karst Aquifer Rocks

There are many kinds of carbonate rocks, and their differences are reflected in their varying capacities to store and transmit underground water. Of particular importance are the characteristics of purity, porosity, fissuring and thickness.

By definition carbonate rocks contain more than 50% carbonate minerals by weight. There are two common end members : calcite ($CaCO_3$) and dolomite ($CaMg(CO_3)_2$). A rock composed of at least 90% calcite is termed a limestone. If its proportion of dolomite increases up to 50% it is termed a dolomitic limestone. More than 90% $CaMg(CO_3)_2$) constitutes dolomite rock.

In New Zealand, dolomite is restricted to the small area of Mt. Burnett near Collingwood in northwest Nelson. But dolomitic limestone is common on some coral islands in the Pacific such as Niue.

Limestones and dolomites may contain insoluble materials such as clay and quartz, and are considered impure if they contain 10-50% of these materials. Rocks with more than 50% impurities are not defined as limestones or even impure limestones but as, for example, calcareous sandstones. Some limestones, such as the argillaceous limestones just north of Auckland, are pure enough to be quarried for agricultural lime, but are much too impure to develop karst. Nevertheless, most New Zealand limestones are more than 90% calcite and are well karstified.

The primary porosity of carbonate rocks varies from about 22% in the Pleistocene coral reefs of Pacific islands to almost zero in the Ordovician Arthur Marble of north-west Nelson. Primary porosity is highest in the youngest rocks, including the Pliocene coquina (shelly) Te Aute limestones of Hawkes Bay and the Wairarapa. In the widespread Oligocene limestones, primary porosity is usually less than 1%, being least where the rock is most crystalline. Hence, the most important karst rocks in New Zealand (the Oligocene limestones and the Arthur Marble) have very low primary porosities. Their water-bearing capacity has depended on the development of a high secondary porosity, although even that rarely exceeds an average of 2%. This may seem very small but, by analogy, in a high rise building with standard plumbing services, the volume of pipes in the entire building is a very small percentage of the total volume of the building. Nevertheless, there is enough water to cause serious flooding if the pipes fracture. So it is in a karstified mountain. Large volumes of water are stored in and pass through a network of relatively small natural pipes. On the other hand, on raised coral islands such as Niue and some of the Cook Islands group, the old reefs are exceedingly porous and karstification is not necessary for a large groundwater body to develop. Nevertheless, because of the solubility of the carbonate rock, large caves can form, and a karst network of passages often directs the flow of underground water from the interior to springs on the coast.

Fissuring of all kinds is extremely important for enabling groundwater flow. This is especially the case where the unweathered rock has very low primary porosity, and bedding-planes, joints and faults provide the main pathways for water movement. Jointing usually increases as bed thickness diminishes. Thus the thinly-bedded, platy Oligocene limestones that constitute most of the outcrops in New Zealand shown on Figure 11.1 are highly fissured and very susceptible to the development of secondary porosity. Consequently they contain many hundreds of caves. By contrast, the very massively bedded Arthur Marble of Nelson contains few bedding-planes and relatively few, widely spaced joints. Cave passages, although large and long, are widely spaced.

The thickness of karst rocks partly determines their potential to store groundwater. The Oligocene limestones are variable in thickness, ranging from a few metres thick near Port Waikato to about 100m around Waitomo. In the South Island these beds can vary from 20-40m near Collingwood and Paturau to over 100m in the Punakaiki syncline between Westport and Greymouth. They are also highly variable in composition throughout this thickness, ranging from very pure and crystalline to sandy, shelly and argillaceous (Figure 11.2). However, even the great Flint Mammoth Cave system of Kentucky is developed in no more than 100m of limestone, so the comparative thinness of a carbonate formation need not unduly restrict the development of a major aquifer. Continuity is also an important factor - the Oligocene limestones of the Te Kuiti Group occur in patches over about 1000km^2 of the western North Island, but few really large aquifers can develop because of their discontinuous distribution. Other patches outcrop on mountain summits, for example the Matiri Tops of the Buller valley; they are freely drained by gravity and so cannot support large underground water bodies.

In contrast with these is the Arthur Marble of

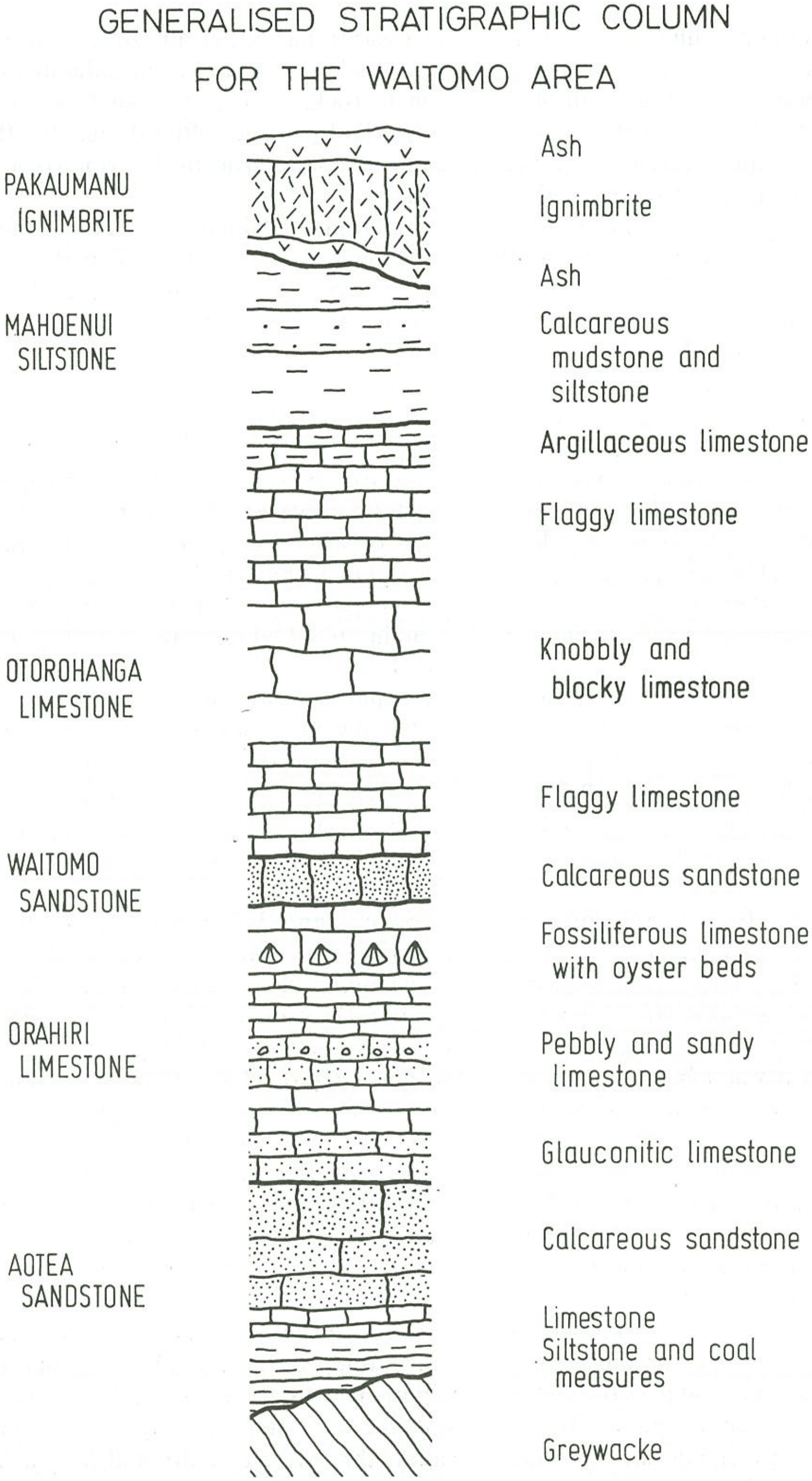

Figure 11.2 Generalised stratigraphic column for the Waitomo area, North Island (From Williams (1992) after Kear & Schofield (1959) and others).

Nelson, which is more than 1 kilometre in stratigraphic thickness and extends in a discontinuous belt for about 90km from Collingwood to Mt. Owen (Figure 11.1). The outcrop is up to 7km wide in places and extends from 1876m above sea level on Mt. Owen to well below sea level in the Takaka valley. In places it contains huge groundwater reservoirs, and it sustains the flow of New Zealand's largest springs - the Waikoropupu Springs near Takaka. Smaller patches of marble of similar age outcrop in Fiordland.

The Functioning of Karst Hydrologic Systems

All the basic concepts of groundwater hydrology introduced in Chapter 10 apply to karst, but some additional factors give karst its special character.

Unlike groundwater systems in other rocks, the circulation of water in a karst aquifer continuously enhances the aquifer's permeability, and thereby progressively modifies the groundwater hydrological system. The capacity of karst rocks to store water increases as the interconnected secondary voids are enlarged by solution. Resistance to water flow within the system simultaneously decreases. Hence, as the storage capacity increases, the water table is lowered and the hydraulic gradient becomes less steep. The water table in karst terrain can consequently lie at a considerable depth below the surface and it can be very nearly flat. For example, the water table beneath Mt. Arthur is controlled by the elevation of the main outflow spring, the resurgence of the Pearse River, and lies up to 900m beneath the upper slopes of the mountain.

Three main hydrographic zones are recognised in karst: (i) the aerated or vadose zone, (ii) the permanently waterlogged or phreatic zone, the top of which is the water table and (iii) an intermediate epiphreatic zone that is periodically flooded as the water table rises and falls. In New Zealand, the epiphreatic zone can be 20m or more in thickness in some karsts. The uppermost, most weathered part of the vadose zone beneath the soil (the subcutaneous zone or epikarst) is typically 10m thick. All of these zones move downwards through the karst as the landscape is lowered by solution. The elevation of the water table is determined by the level of the outflow spring that drains the aquifer. Hence, as the spring is lowered by downcutting of the river into which it flows, so is the water table. In Nettlebed Cave beneath Mt. Arthur, the water table has been lowered at a rate of 440cm per 1000 years (Williams 1987).

Recharge in Karst Systems

A fundamental consideration in understanding how a karst groundwater system functions and has developed is the origin of the waters that recharge it. A distinction should be made between the precipitation that falls directly onto the limestone outcrop (termed autogenic recharge) and that which falls on neighbouring but non-carbonate rocks, from which it flows as streams or rivers onto the karst (termed allogenic recharge). These two forms of recharge have major implications for the aquifer. Precipitation falling onto karst is dispersed relatively evenly and has a distinctive chemistry. This type of recharge and associated corrosion results in the development of an extremely intricate network of interconnected fissures and pores, especially in the subcutaneous zone beneath the soil. Hence it is responsible for the generation of most of the diffuse near-surface storage in the groundwater system. By contrast, allogenic recharge is already collected into streams when it flows onto the karst, consequently recharge is focused at the place where streamflow is lost underground. The chemistry of streams flowing onto karst reflect the terrain where they originated, and their erosional capacity may differ from that of rainwater. Incoming streams commonly disappear underground suddenly at stream-sinks, although some streams lose their flow gradually by seepage into their beds. Sinking streams flow through cave passages to springs and are mainly responsible for the development of caves.

Geology and topography determine whether autogenic or allogenic recharge will dominate the development of a karst hydrologic systrem. Isolated karst plateaus with little or no impermeable coverbeds, such as Mt. Owen in northwest Nelson, Mt. Kahurangi on the East Coast, or uplifted coral reefs such as Niue Island, are mainly recharged diffusely by precipitation. But where higher impervious rocks drain onto karst or where dissected, impervious coverbeds shed water onto underlying limestones, then allogenic recharge results. This is a most common occurrence and is found throughout the King Country where Mahoenui siltstones drain onto the limestones beneath, in the Abel Tasman National Park where streams collected on high granite inliers flow onto surrounding but lower lying marble, and in some of the Cook Islands where water is shed radially from interior volcanic uplands towards a ring of uplifted coral reefs. Nevertheless, in practice most karsts are recharged both by direct rainfall and streamflow.

Storage and Transfers in Karst Hydrologic Systems

Karst groundwater systems can be visualised as a series of stores of water, such as the subcutaneous store and the saturated (phreatic) zone store, interconnected by passages through which water flows until it reappears at a spring (Figure 11.3). Some stores, such as the subcutaneous zone store, are small and can hold comparatively little water

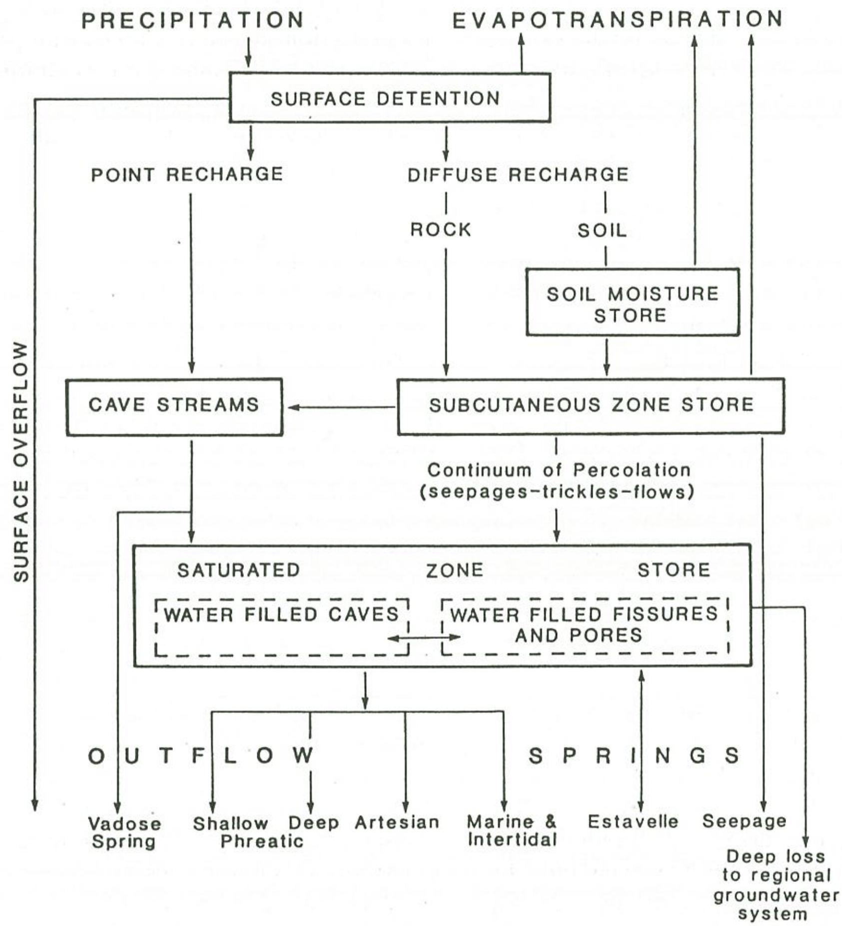

Figure 11.3 Stores and linkages in karst hydrologic systems. The soil moisture store, subcutaneous zone store and cave streams are located in the vadose zone. The saturated zone store is in the phreatic zone. (From Ford & Williams 1989)

in them at any one time, but transmit a large volume during the course of a year, whereas others can be very large, such as some phreatic stores, but have relatively little water passing through them. In the first case, the turnover time of water in storage is short, perhaps a few weeks, whereas in the second it may be many months or even a few years.

The connections between the stores are fissures, pipes and pores of various sizes. Some of the largest are caves many square metres in cross-sectional area through which underground rivers flow. In the vadose zone these rivers are similar to those at the surface, except that they have a roof, but within the phreatic zone they completely fill the cave, have no free surface and may flow under pressure. In the phreatic zone, the interstices of the rock are full of water derived both from vertical percolation from the vadose zone and from the back-flooding of phreatic rivers during high discharge events. There is a net movement of water from these fissures into the flooded cave passages and thence to the spring.

Darcy's Law (Chapter 10) describes groundwater movement in the phreatic zone, below the water-table, **provided** the movement is slow enough to be laminar. In karst this is usually not the case, because water flow in pipes and caves is frequently turbulent (Ford & Williams, 1989). Nevertheless, in some areas of New Zealand and the South Pacific, Darcy's Law can be applied to karst. Important cases include, for example the large groundwater body in thick marble beneath the Takaka valley, and the uplifted coral islands of Niue, the Cook group, Tonga. Groundwater flow through karst, therefore, tends to vary between the extremes of turbulent flow through pipes on the one hand and Darcian flow through pores and fissures on the other, with most karsts having a mixture of both (Figure 11.4).

Because karst hydrologic systems are largely subterranean, they are difficult to observe and study. Nevertheless, a great deal can be deduced about the underground system sustaining the flow of a spring from observation of the spring itself. The karst groundwater system is virtually a "black box", but when it rains the recharge stimulates a response at the spring. The nature of the response depends on the characteristics of the groundwater system it has passed through. Typically two simultaneous signals are measured at the spring : the variation in its discharge and the variation in the chemical quality of its water. Electrical conductivity is often used as an index of mineralization of

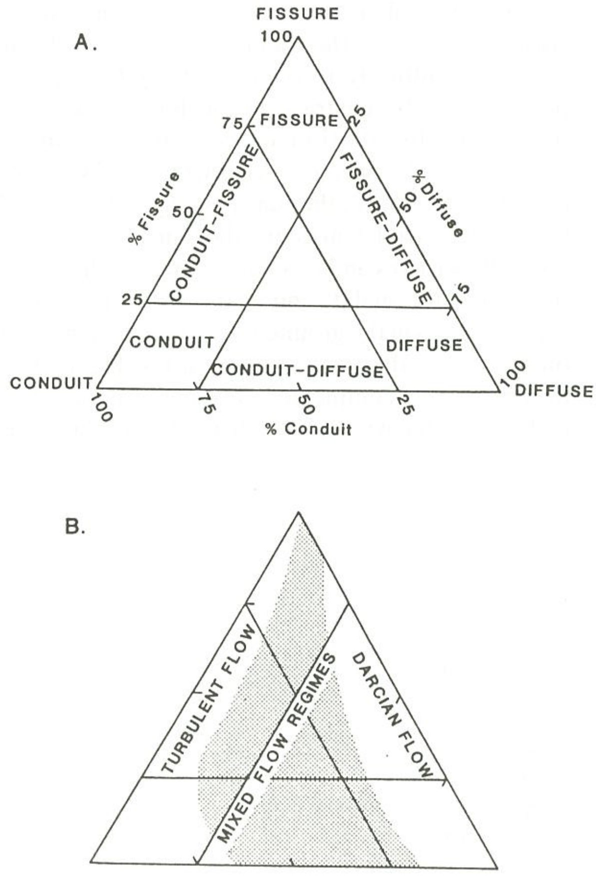

Figure 11.4 A conceptual classification of karst aquifers and their presumed relationship to predominant flow regimes. In **4A** a distinction is made between carbonate aquifers dominated by diffuse pattern of flow through interconnected pores, as opposed to strongly directed flow through caves (conduits), or flow through a geometrical network of fissures (joints). As indicated in **4B**, these styles strongly influence whether the groundwater flow will be dominantly Darcian (laminar), turbulent, or a combination of both in different parts of the aquifer .

the water. The responses to rainfall of both discharge and conductivity at a spring are rapid when recharge is dominated by sinking streams and groundwater flow passes mainly through caves. But the responses can be very lagged when recharge is dispersed and groundwater flow is Darcian. In the latter case, the frequency distribution of recorded conductivity values is typically unimodal, because of the similar chemical history of the water that infiltrates slowly and flows diffusely through a highly fissured or porous rock. By contrast, in the former case, the frequency distribution of conductivity values recorded at the spring is characteristically multimodal (Figure 11.5), the reason being that in a well karstified, mixed autogenic/allogenic system, different flow routes and residence times result in contrasts in water quality, and as the water from these various places in the groundwater system emerges at the spring, so the water quality varies. Figure 11.3 illustrates the complexity of some systems. The highest conductivity peak is from water which has passed through the subcutaneous zone, where it dissolved substantial amounts of limestone; the lowest conductivity peak is the chemical "fingerprint" of relatively unmineralized, rapidly flowing conduit (cave) water from stream-sinks; and the intermediate peaks are from water with a moderately low residence time that has passed through widened fissures. Most karsts in New Zealand are of this well developed complex type.

Water Tracing and Pollution Tracing

Springs are traditional sources of drinking water. People tend to assume that soil and rock filter and clean water, so that when it re-emerges it is essentially free of bacterial and chemical pollutants and is safe to drink. In fact, karst is an extremely bad filter, and in agricultural countries like New Zealand karst spring water often is not safe to drink without boiling.

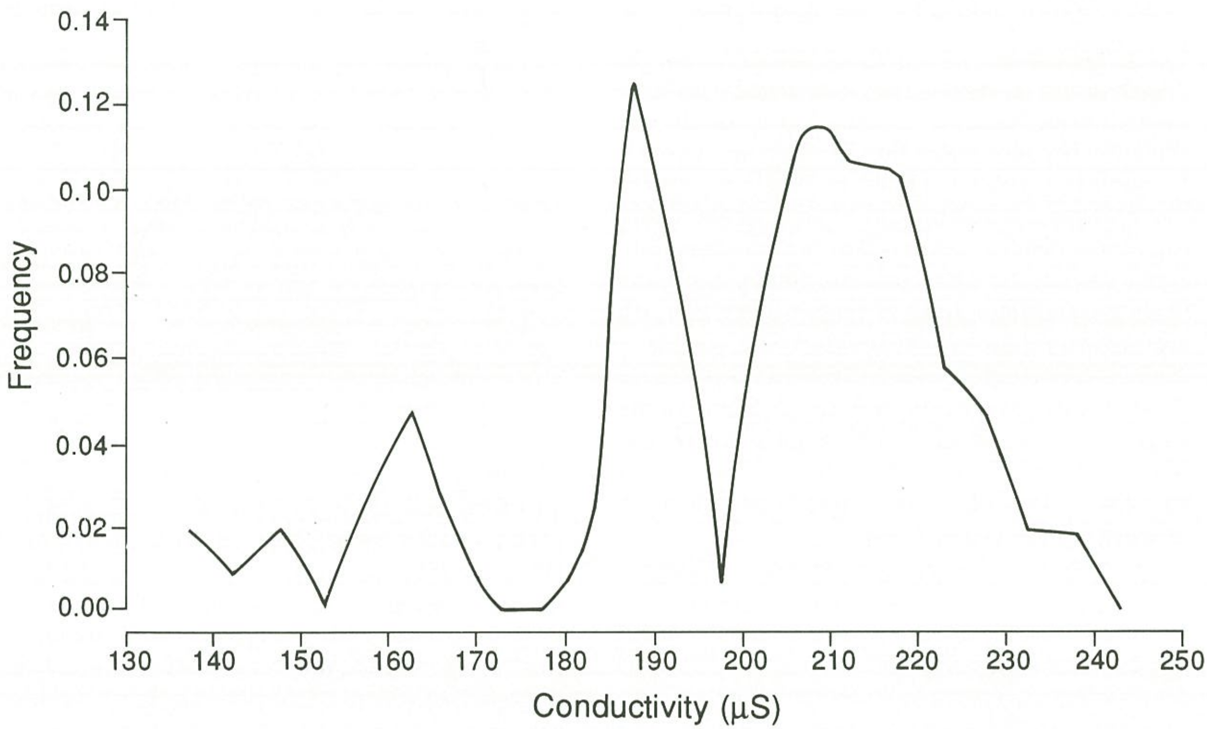

Figure 11.5 Frequency distribution of electrical conductivity values from a karst spring flowing from a limestone catchment at Mangakowhai, near Te Kuiti. The multimodal nature of the curve is typical of well karstified aquifers with mixed flow regimes (see Figure 4B). (From K Wood, unpublished)

How many tomos (sinkholes) used for rubbish disposal are there on farms in our karst areas? Probably every farm on karst, not just in New Zealand but all round the world, has used naturally occurring sinkholes as convenient places to dump dead stock or domestic refuse etc. Out- of-sight out-of-mind is a common philosophy. Yet literally thousands of water-tracing tests by karst hydrologists have shown flow velocities from sink to spring to be typically in the range of 5-100 metres/hour. In other words, farm refuse washed by rain from a tomo into an underground stream can easily travel, say, a kilometre to a neighbour's spring in a day or so. This is insufficient time to kill bacteria, since most species of bacteria require more than 100 days to die. Physical filtration in karst conduits is also negligible.

Because karst springs in many parts of the world are very important water supplies, considerable effort has been made to develop techniques to trace the source of their waters, and to define the catchment areas from which they are recharged. When the catchment has been defined, land-use practices can be managed to protect the quality of groundwater.

By far the most effective way of tracing underground waters is by means of fluorescent dyes. The most commonly used dye is fluorescein ($C_{20}H_{12}O_5$), a green dye that was first used as a tracer of water sinking into the limestone bed of the upper River Danube in 1877. This dye is ecologically safe, but is less efficient than Rhodamine WT. The latter is a modern orange dye designed especially for water tracing, that under good conditions is detectable in concentrations of as little as 1 part per billion. These and other fluorescent dyes can be detected using instruments such as fluorometers and UV spectrophotometers (Ford & Williams, 1989).

Major Karst Aquifers

The Takaka Valley

The most important karst aquifer in New Zealand in terms of the volume of water in storage is in the Takaka valley, northwest Nelson (Williams 1977, Stewart & Williams 1981, Stewart & Downes 1982, Mueller 1992). It is found in Arthur Marble of Ordovician age that underlies the north-south fault-angle depression of the Takaka valley and outcrops on the surrounding hills to the east and west (Figure 11.6). The marble extends over about $180km^2$ and has a stratigraphic thickness of more than 1000m, so its water storage potential is considerable. The downstream $45km^2$ of the marble is capped with Tertiary coal measures, so the outflow zone of the aquifer is confined under artesian conditions.

Three major rivers drain from surrounding highlands into the valley: the upper Takaka river, the Waingaro and the Anatoki. Each river loses flow into its bed when it crosses the marble, providing a major source of recharge (Figure 11.7). The upper Takaka river has an average flow of $16.1m^3/s$, but loses about $11m^3/s$ into the aquifer by percolation into the fluvial gravel which veneers the marble. Since the flow of the river is less than 11 m^3/s for about 100 days per year, the river bed across the marble is dry for long periods of time. However, the similar sized Waingaro and Anatoki rivers together only lose about $2m^3/s$ of their combined flow, because they cross narrower strips of marble. Further allogenic recharge comes from numerous small streams which drain into the marble from higher non-karst rocks in the surrounding uplands.

Autogenic recharge is contributed by rain falling on the $90km^2$ of marble outcropping on the surrounding hills, much of it within the Abel Tasman National Park. Further recharge comes from rain falling onto $45km^2$ of gravel flats overlying marble in the central portion of the valley floor. Given evapotranspiration losses of about 700mm, annual recharge in these areas totals about 1900mm, or approximately $8m^3/s$.

The total annual recharge to the marble aquifer is estimated to be about $23.4m^3/s$ (Mueller, 1987,1992). The main outflow from the aquifer, the Waikoropupu Springs (Figure 11.8), has a combined discharge of $15m^3/s$, leaving a difference of $8.4m^3/s$ to be accounted for. Submarine springs are known to exist on the bed of Golden Bay, so one may presume that this unaccounted sum is their annual

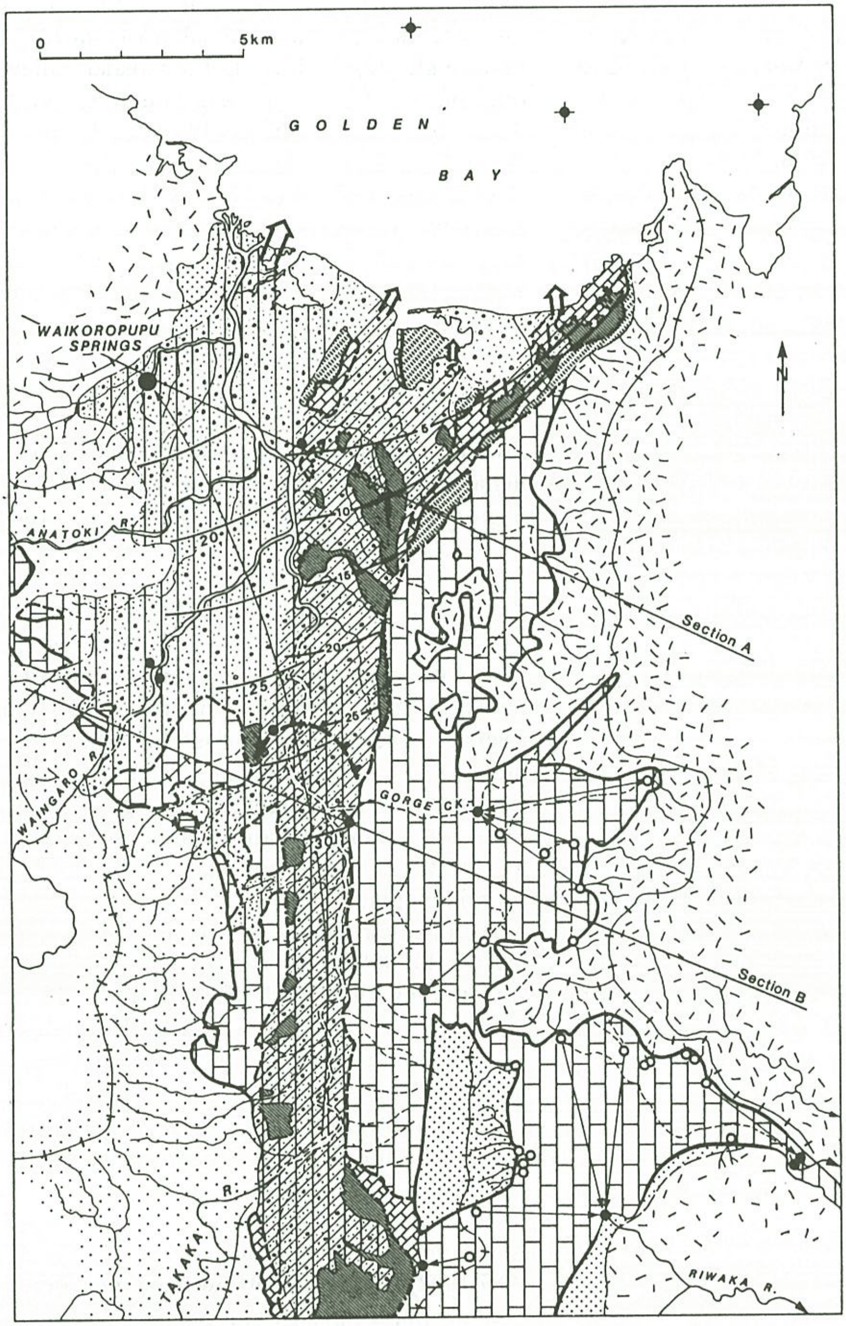

Figure 11.6 A hydrogeologic map and cross-sections of the Takaka valley. (From Ford & Williams 1989)

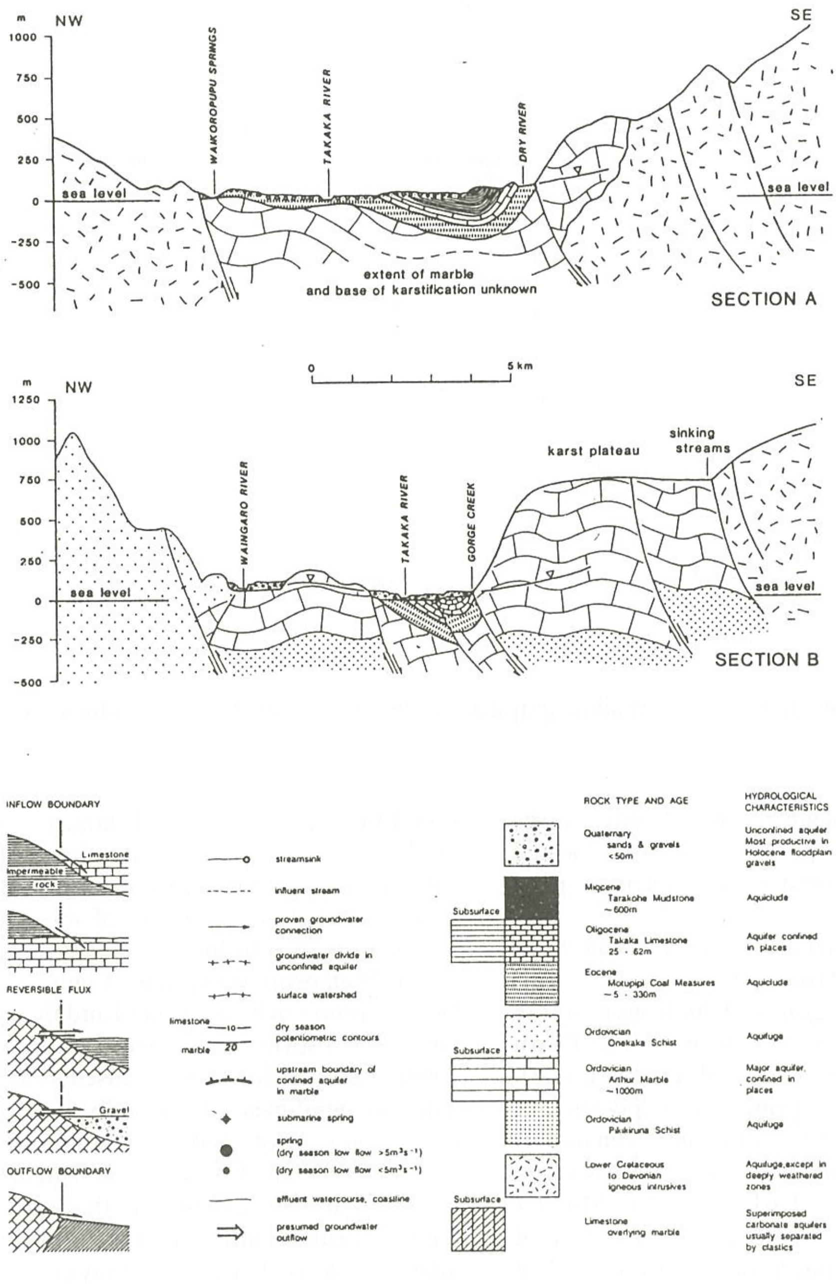

Figure 11.6 (cont) A hydrogeologic map and cross-sections of the Takaka valley. (From Ford & Williams 1989)

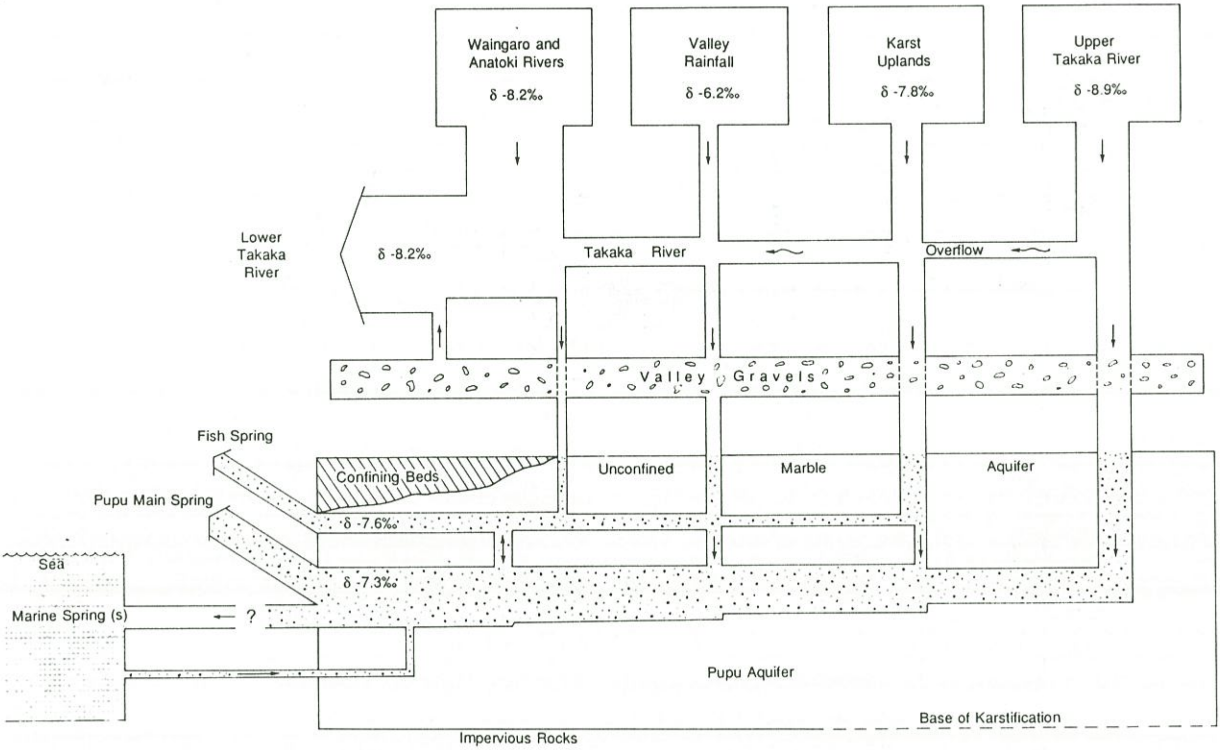

Figure 11.7 A schematic model of the Waikoropupu groundwater system. Delta (δ) values refer to oxygen isotope $\delta^{18}O$.

outflow. However, despite recent scientific efforts to find these springs, they have yet to be located and their flows measured (A.D. Fenemor, pers. comm.)

Dating of Pupu Springs' water by tritium (H^3) and characterisation of recharge and groundwaters by oxygen and hydrogen stable isotope analyses have shown the Pupu hydrogeologic system to be complex (Figure 11.7). Water emerging at the springs has a spectrum of ages, being a mixture of water that has been underground for various lengths of time and has come from different sources. Taylor (cited in Mueller 1992) considers there to be a base reservoir of water that has been underground for longer than 10 years, but not more than 20 years, supplemented by a much more recent component that Stewart (cited in Mueller 1992) identifies to be about one year old. Although Stewart & Williams (1981) found the aquifer to be well mixed, it cannot be modelled as a simple exponential reservoir, because it also contains elements of piston flow. It is best envisaged as a cascade of mixing compartments connected by linear conduits.

The Waikoropupu Springs (Figure 11.8) are the largest springs in New Zealand and the 24th largest known karst springs in the world (Ford & Williams 1989). Their ecology is recognised as having international significance (Michaelis 1976, 1977) and they are now protected as a scenic reserve. This gives rise to some difficult management problems, because water extraction from the aquifer sustaining the springs will reduce their flow and human activity in the recharge zones may contaminate the groundwater and its ecosystem. This problem is being addressed by the Nelson-Marlborough Regional Council in its "Takaka Valley Water Management Plan".

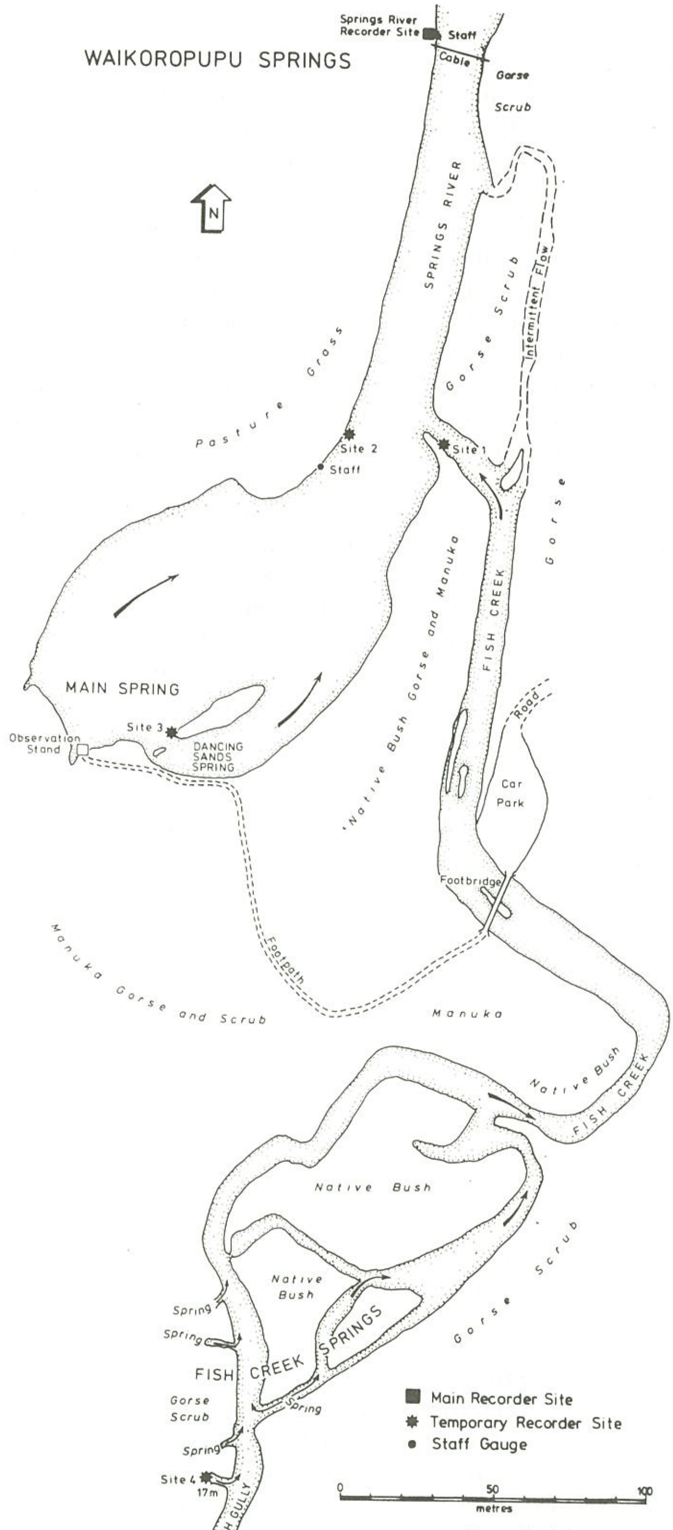

Figure 11.8 The Waikoropupu Springs and Springs River near Takaka. (From Williams 1977 after Rapier 1975)

A dilemma arises because of water shortage in the face of plenty. The outflow from the Waikoropupu Springs to the Springs River is $15m^3/s$ on average with a maximum of about $21m^3/s$. The time of travel through the system varies widely, but has an average of roughly 3 to 8 years; hence the volume of storage in the underground reservoir must be considerable, and at least $1.5km^3$, which should be sufficient water for local needs. However, the total amount of storage is less important for water supply allocation than are the dynamic reserves, i.e. the volume involved between the extremes of high water outflow and low water outflow. Since the lowest flow so far measured is approximately $5.3m^3/s$, and agricultural demand for water is often greatest during droughts, too much pumping from the aquifer or abstraction from the spring may seriously endanger the Springs River ecosystem. Because of existing water rights for salmon farming beside the Springs, the Regional Council has set an interim limit of $0.5m^3/s$ for total abstractions from the recharge zone for the Waikoropupu Springs, and the minimum residual flow in the Springs River is set at $2.9m^3/s$ (20% of the natural mean discharge). Maintenance of water quality places further constraints, as leaching of materials from agricultural chemicals and stock in the recharge zone and waste water contaminants from the salmon farm may cause pollution.

Niue Island

The aquifers found in uplifted coral reefs of such islands as Niue, Tonga and the Cook Islands group contrast strongly with karst aquifers developed in Arthur Marble. They have a much higher primary porosity, but a lower secondary porosity than those in the Arthur Marble.

As Niue Island consists only of coral, it is an ideal autogenic system, being recharged entirely by rainfall falling directly onto the carbonate outcrop. It has a comparatively simple, roughly elliptical shape and is built on a base of volcanic rocks located 300m below sea-level. It is an emerged atoll with a former encircling reef and interior lagoon uplifted to about 60m. The island has an area of $258km^2$, and is somewhat lower in the interior on what was the former lagoon floor. The predominantly dolomitic coral of which the island is composed is late Tertiary to Pleistocene in age.

Niue illustrates very well the way in which fresh water, with a of density of about 1.00 $g.cm^{-3}$, floats on salt water with a density of 1.025 $g.cm^{-3}$. Because of the difference in densities, for every 1 metre that the fresh water-table stands above sea-level, the depth to the underlying fresh water/salt water interface is 40m, provided the situation is hydrostatic, i.e. there is no circulation. However, recharge of most natural hydrological systems, generates flow and circulation, so that the depth to the fresh water/salt water interface is rather more than 40 times the height of the fresh water above sea-level.

This phenomenon was first investigated at the turn of this century by two European scientists, Ghyben and Herzberg, whose names are now lent to the principle that they discovered. The important implication is that, if the water-table rises in a gentle dome towards the interior of an island, then the fresh/salt water interface will descend steeply downwards. It will reach its greatest depth beneath the place where the water-table is highest, which is usually (but not always) the middle of the island. The fresh water body in an island is therefore usually lens shaped.

Niue Island has many boreholes drilled to extract groundwater and many caves, especially near the coast, which can be explored down to the water-table. This has allowed measurement of the elevation of the water-table above sea-level at 25 locations (Figure 11.9). The shape of the water-table is not a simple dome rising towards the centre of the island. Instead it consists of a discontinuous string of elliptical domes around the outside of the island, beneath the highest land that was once the atoll that enclosed the lagoon. The water-table attains 3m or so above sea-level on these domes, whereas in the centre of the island it reaches only 1.6m above sea-level. There are low spots between the encircling domes that provide avenues for water from the interior to escape to the coast. Springs in the inter-tidal zone on the coast sometimes are located opposite these gaps (Figure 11.9).

As a result of the topography of the water-table, the fresh/salt water interface at depth should have

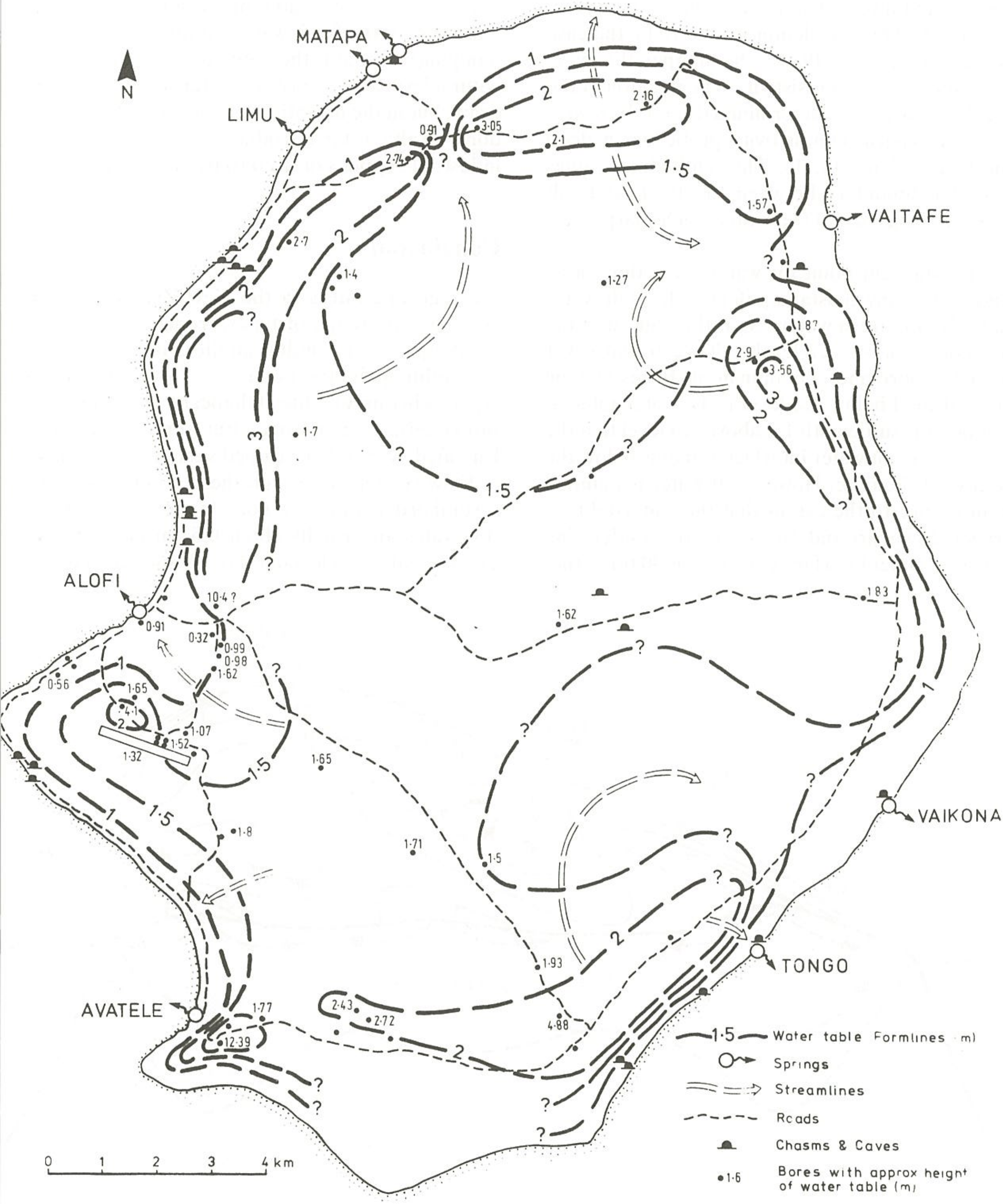

Figure 11.9 A water-table map for Niue Island, south Pacific. In most locations the water-table lies about 60m below the surface of the island.

a similar but inverted and exaggerated shape (Figure 11.10). This was demonstrated to be the case by Jacobson & Hill (1980), who mapped the interface using electrical resistivity. The position of the interface can also be determined directly by lowering an electrical conductivity probe down deep boreholes (Figure 11.11). The conductivity values show the boundary between the fresh and salt waters to be a zone of transition, not a sharp interface.

An important point for water extraction on islands (or in any coastal aquifer) is that salt water underlies the fresh water. So if the bottom of the borehole is above sea-level, salt water can never enter the bore and contaminate it. Bores may be safely drilled below sea-level if the water-table at that point is sufficiently far above sea level that the fresh water/salt water interface remains below the bottom of the bore. However, if water is pumped from a bore to the extent that the water-table is drawn down around the bore, the underlying fresh/salt water interface must rise by 40 times that fall. As a result, sea-water may invade the bore and contaminate the fresh water supply. Cessation of pumping will enable the water-table to return to its natural level and so force the interface downwards again, but in the meantime damage may have been done if salty water introduced into domestic and industrial supplies or used to irrigate crops.

Conclusion

Karst groundwaters in the New Zealand region cover a wide range of types, from highly porous coral aquifers to conduit aquifers in marble with negligible primary porosity. The widespread aquifers in thinly bedded Oligocene limestones are intermediate in nature, but tend to be well karstified, with well developed systems of conduits.

Flow velocities through these aquifers vary by several orders of magnitude. In the coral of Niue, flow rates are usually about 0.5-1m per day, but exceptionally reach 300m/day. A similar range of

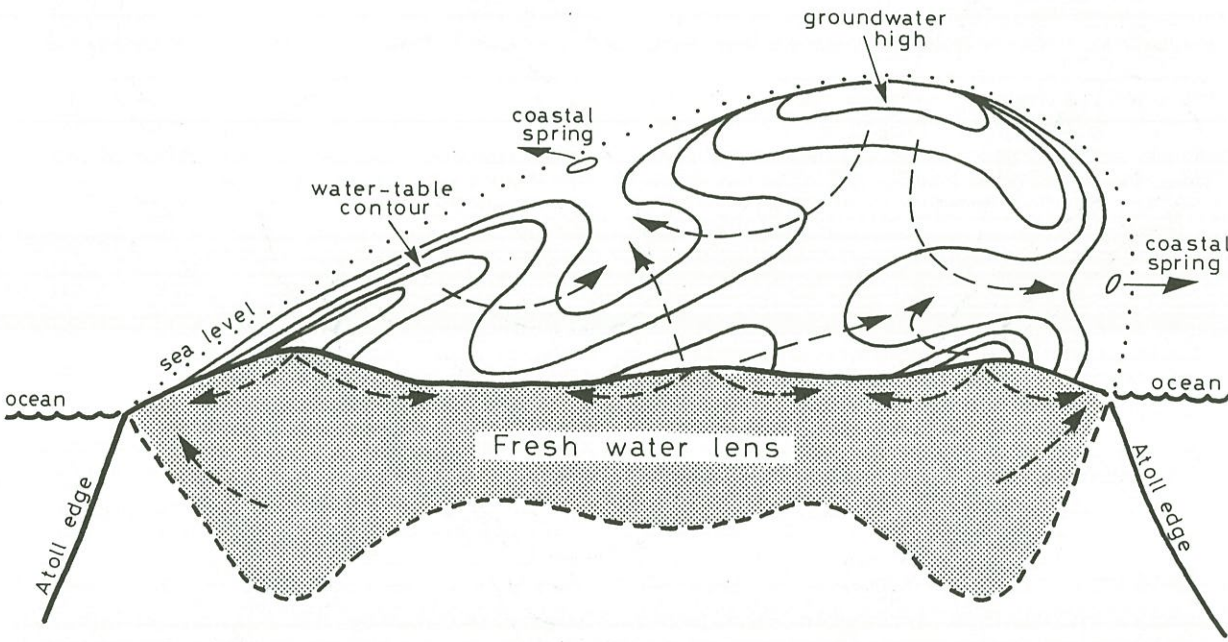

Figure 11.10 A block diagram of the form of the fresh water lens beneath Niue Island. Note the relationship of the topography on the water-table to the depth of the fresh water/salt water interface.

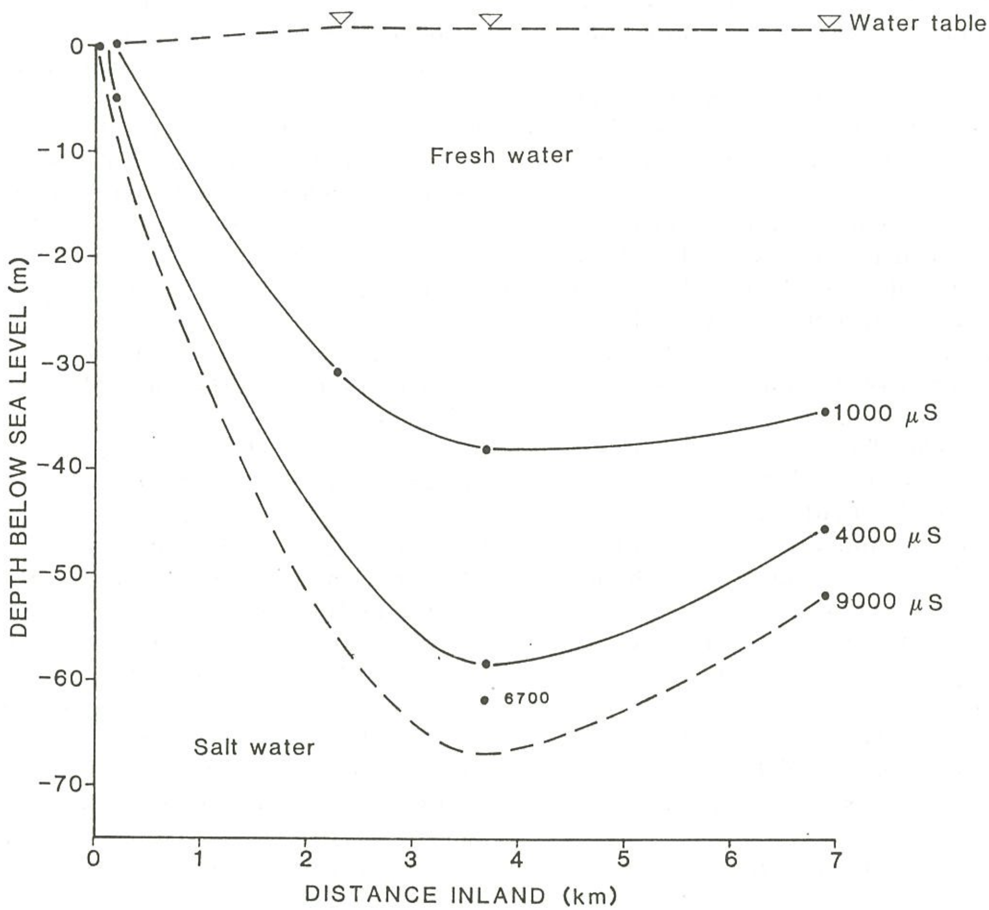

Figure 11.11 The location of the fresh water/salt water interface beneath Niue Island as determined by electrical conductivity sounding of boreholes.

flow rates occurs in the Waikoropupu aquifer of the Takaka valley. But in areas of high relief where there is greater fall from the recharge zone to springs, flow rates can be much faster, exceeding 1km/day in the marble karsts of Mts. Arthur and Owen and in the Oligocene limestone karsts of the Waitomo district or the Paparoa National Park.

Rapid transit times associated with recharge by sinking streams introduces pollutants very speedily into karst aquifers, which are therefore very vulnerable to pollution. The large size of karst conduits also means that physical filtration is minimal. Hence although karst springs may look attractive, there are few in New Zealand that are safe to drink without first boiling the water.

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12
Water Quality in NZ

R A Hoare and L K Rowe

"Pure" Water.

The popular conception of water in its natural state in rivers, lakes and groundwater is that it is "pure". In some mystical sense, this may be so. However, in the scientific sense this is very far from the truth. One of the reasons water is so important to the environment and to living things is that it is a very good solvent, and it almost always contains dissolved impurities. These impurities provide nutrition to plants, and they even supply some vital trace elements to animals and to people. ("Pure" water is the main source of dietary fluoride to people, even when none is added artificially.)

"Quality" is a very difficult concept to define, even though most people have some idea of what they mean by it. (Pirsig, 1974.) Water "quality" usually comes to our attention when the level of impurities becomes too high for us to be able to use the water for specific purposes. However, Maori culture defines the quality in terms of what is **done** to the water, rather than what is **in** it. We will describe quality in quantitative, technical terms in this chapter. The focus is on the physical and chemical aspects of water quality, rather than on the micro-biological or biological aspects.

This chapter will describes the kinds and concentrations of solutes one finds in fresh water in New Zealand, their sources, how to estimate the amount of solutes coming from these sources, how to estimate the effect of the loads of material to surface water, and how to measure the water quality. It concludes with a discussion of the results of a survey of the Regional Councils, in which opinions on current water quality were sought.

The Maori Perspective

In New Zealand in the 1990's it is essential to take cognisance of Maori attitudes to environmental matters. The authors have sought the views of the Tainui in this matter, since they have been in the forefront of the discussions and legal hearings which are presently establishing the role of these views in environmental management.

We were referred to the Chilwell decision of the High Court, (12 NZTPA, P 129) and to an unpublished report of the Huakina Development Trust to the Waikato Regional Council on the Waikato Regional Coastal Plan, for statements on the Maori attitude to water quality. These reports were specific to situations being discussed, but the following quotes from the report to the WRC are relevant to this chapter.

(Page 30) "According to Maori tradition even water has a mauri (spiritual life force). The mauri is the force that ensures within a physical entity such as the sea, harbours, rivers, lakes and estuaries including land, that all species that it accommodates will have continual life. The mauri cannot be interrupted or desecrated. However, if it is, whatever it accommodates is at risk.

The Tangatawhenua believe that disasters or natural phenomena cannot harm the mauri, only that instigated by man, merely by the use of artificial components such as chemicals. The mauri is defenceless against components that are not part of the natural environment. The mauri of waters and the wairua (spirit) of the Tangatawhenua have the same origin. Therefore when the mauri is harmed, so too is the spirit of the Tangatawhenua."

(Page 66) "All direct discharges of waste should be prohibited into waterways ... without first being treated and purified by passing through the land."

It is apparent from these quotations that the Maori attitude to water quality is not focused on the constituents of the water, so much as on the processes that the water has been subject to. In the High Court decision, the Waitangi Tribunal is quoted as saying "Waste water is purified by return to the earth, ritualistic purification, or, with the exception of water containing animal wastes, by mixing with large quantities of other pure water."

In our view, the chemist and the Maori are looking at different things when they describe water quality. Water that is chemically harmless can offend Maori sensibilities, and water that has been passed through soil and therefore satisfies the Maori view on "purity" can contain a number of constituents that are chemically undesirable (notably, nitrate). It is up to the potential user of water to decide whether it is appropriate to assess its quality from a Maori or a chemical viewpoint.

Substances Commonly Found in Water

Major Ions.

These are ions such as sodium, potassium, calcium, magnesium, chloride, silica, bicarbonate and sulphate which are present in all waters in concentrations from about 1 $g.m^{-3}$ upwards. New Zealand waters are generally considered to be of very high quality by worldwide standards. A major study of 100 New Zealand rivers (Close and Davies-Colley, 1990) showed median concentrations of the major ions (with the exception of sodium and silica) to be lower than the worldwide average. Ion concentrations in United States waters were much higher than in New Zealand ones. Streamwater ionic concentrations are presented in Figure 12.1[1] (data from Close and Davies-Colley, 1990). They have a wide range, with high values at least 10 times, and sometimes 50 times, greater than low values. The upper and lower quartiles usually differ by a factor of about 3. The concentrations are approximately log-normally distributed, and this needs to be taken into account when using statistical tests to compare concentrations at different sites.

Suspended Solids

Suspended solids concentrations in the "base flow" samples of the 100 rivers project (ie, samples taken when the flow rate was less than the median flow rate of the stream) had lower and upper quartiles of 0.9 and 5 $g.m^{-3}$. (Note that these are less than those of the dissolved constituents.) In the past, suspended solids concentrations have been used as an important measure of water quality. This measure still has use for estimating rates of

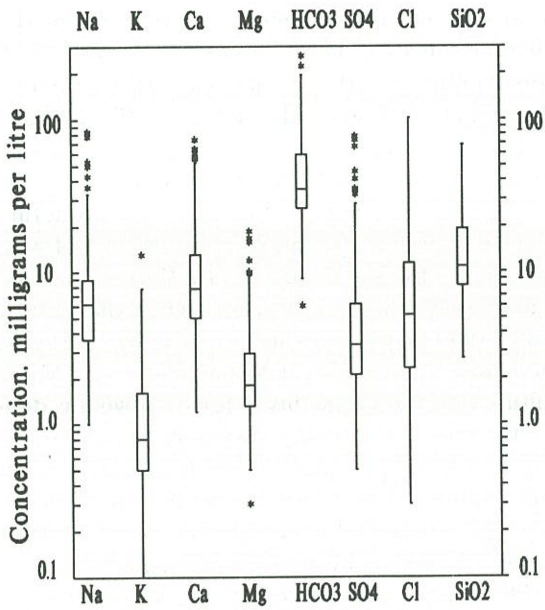

Figure 12.1 Concentrations of major ions found in the 100 rivers study.

[1] This figure presents box plots of the data. The 3 horizontal lines represent the upper quartile, the median and the lower quartile. The vertical line represents the range, unless there are "outliers", shown as asterisks. See a modern statistics book for the definition of an outllier.

erosion, and as an easy way to describe an aspect of water quality. However, recent work in NZ and Australia has shown that the important effects of suspended solids are optical ones, and that there are advantages to be had from measuring the optical effects directly (Davies-Colley, 1991). (See later discussion on physical properties.)

The scattering of light by the sample (the turbidity) is often related to the concentration of suspended solids, for any particular type and size range of solid. This turbidity is measured in a "nephelometer", which is standardised with respect to a substance called formazin. The resulting units of turbidity are called Nephelometric Turbidity Units (NTU) or Formazin Turbidity Units (FTU).

Turbidities are normally of the order of a few NTU in uncontaminated water, when suspended solids concentrations are of the order of a few $g.m^{-3}$. There is often a roughly linear relationship between the turbidity and the suspended solids, but the actual form of the relationship will depend on the particle size distribution and density of the suspended solids.

pH.

This measurement is related to the hydrogen ion concentration. Water can split into hydrogen ions and hydroxyl ions, and the product of the concentrations of the two species is constant. At low pH, high concentrations of hydrogen ions are present, and the water is acid. At high pH, the hydrogen ion concentration is low, the hydroxyl concentration is high, and the water is alkaline. New Zealand streams are generally slightly alkaline with most streams having a pH between 7.5 and 8.5 (Close and Davies-Colley, 1990). Some streams draining forests and wetlands in Westland have pH values as low as 4.1, because they contain high concentrations of naturally occurring organic acids. New Zealand does not have streamwater acidification brought about by high concentrations of sulphate and nitrate in acid rain (Stenzel and Herrmann, 1990), which can be a consequence of the industries and high vehicle density found in many parts of Europe and the United States.

Hardness

Hardness is a measure of the concentration of calcium and magnesium salts in the water. It is very important from an industrial point of view, but from a water quality viewpoint its importance mostly comes from the fact that high hardness helps to reduce the toxicity of many metallic species. High hardness causes a scum when you use soap, but it is not toxic.

Oxygen.

This gas dissolves in water from the air. It can also be released into the water from aquatic plants, when light is available, since it is a by-product of photosynthesis. Oxygen is lost from water by being consumed by organisms such as bacteria or plants. In water with high concentrations of plants, the oxygen concentration can be very high in the day time, and very low at night.

There is an equilibrium between oxygen in the air and the oxygen dissolved in clean water, the concentration in water at equilibrium (or "saturation") being mainly (and strongly) dependent on temperature. Between 1 degree Celsius and 39 degrees Celsius the dissolved oxygen concentration in air-saturated fresh water (in $g.m^{-3}$) is expressible as:

$$DO_{sat} = 14.4 - 0.338*temperature + 0.00354*(temperature)^2$$

with an error no greater than 0.1 $g.m^{-3}$ (Figure 12.2).

It is sometimes useful to express dissolved oxygen measurements in terms of "percent saturation", or the actual measurement divided by the equilibrium value, times 100. A saturation less than, say 60%, means that some process is consuming oxygen much faster than it can be replaced by diffusion into the water, or that there has been a rapid temperature decrease. It usually indicates the presence of organic contamination.

A saturation greater than 100% can indicate the effect of aquatic plants in daylight. It may be associated with very low saturations at night, since

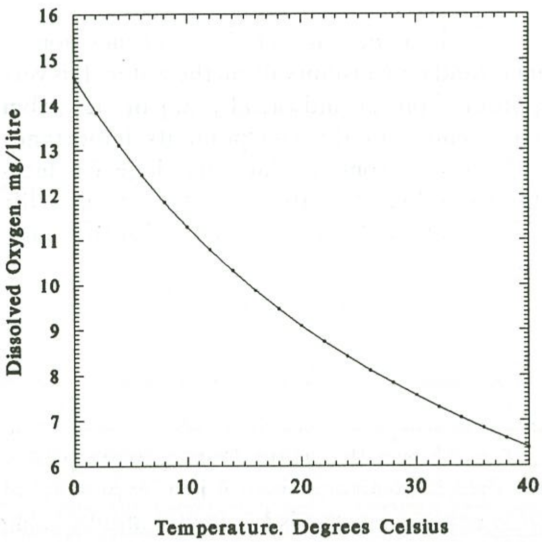

Fig. 12.2 Variation of dissolved oxygen with temperature in air-saturated water.

the plants are consuming oxygen by respiration all the time, but are only photosynthesising in daylight hours.

For healthy fish populations of most species found in New Zealand, a dissolved oxygen concentration of more than 5 $g.m^{-3}$ is desirable. For spawning of salmonids, near-saturation is required.

Nutrients

These are substances that aquatic plants need to sustain their growth. There have been arguments in the past as to whether bicarbonate or minor elements should be included in this category, but in NZ the usage is restricted to the various forms of phosphorus and nitrogen whose absence can be shown to restrict plant growth in our waters. The environmental chemistry of phosphorus and nitrogen is extremely complex, and will only be described in general terms here.

Concentrations of nitrogen and phosphorus are generally expressed as $mg.m^{-3}$ of the element N or P. In some circles, though, nitrate is expressed as the concentration of the compound NO_3, which is about 4.4 times the concentration of N. Similarly, 1 $mg.m^{-3}$ phosphorus as P is equivalent to 3 $mg.m^{-3}$ as PO_4. Be especially careful when reading the older literature. This chapter always expresses the concentration in $mg.m^{-3}$ of P or N.

Phosphorus. There are many forms of phosphorus in water. Often, these are loosely referred to as "phosphates".

Some of the phosphorus forms are loosely defined by the analyses used to determine them (Table 12.1).

P form	Analyzed by:
Available P (DRP)	Filter the water (pore size less than 1 micron). Analyse.
Total P (TP)	"Digest" all the dissolved and organic phosphorus, then analyse
Total Dissolved P (TDP)	Digesting a filtrate before analysing.
Particulate P	TP-TDP
Dissolved organic phosphorus (DOP)	TDP-DRP

Table 12.1 Forms of Phosphorus in water

The DRP is a dissolved form, likely to be immediately available to plants.

TP is not a true total, since it usually does not include all the phosphorus incorporated into inorganic particulate material. However, it includes all that phosphorus that is likely to become available to support the growth of plants through short to medium term chemical or biological processes.

Because of the many forms of phosphorus in water, and the many minor variations in the way that phosphorus is measured, it is vital that when data on this element is gathered with a view to long term trend surveys, the details of the method actually used are stored with the data. Experience has shown that old data is frequently not able to be compared with recent data because of uncertainty about the methods used.

Natural levels of phosphorus in water in NZ are frequently below 10 mg.m^{-3}, and sometimes below 1 mg.m^{-3}. (Figure 12.3.) It is very difficult to reliably measure phosphorus at these low concentrations, particularly if the laboratory is also trying to measure sewage P at 5000 mg.m^{-3}. Much data gathered before 1975 is unreliable at these concentrations. Only exceptionally can one rely on present data below 2 mg.m^{-3}, although some laboratories do an excellent job well below this.

Figure 12.3, based on the data of Close and Davies-Colley (1990), shows the distribution of DRP and TP concentrations found in the 100 rivers study. Median DRP was about 5 mg.m^{-3}, while median TP was about 17 mg.m^{-3}.

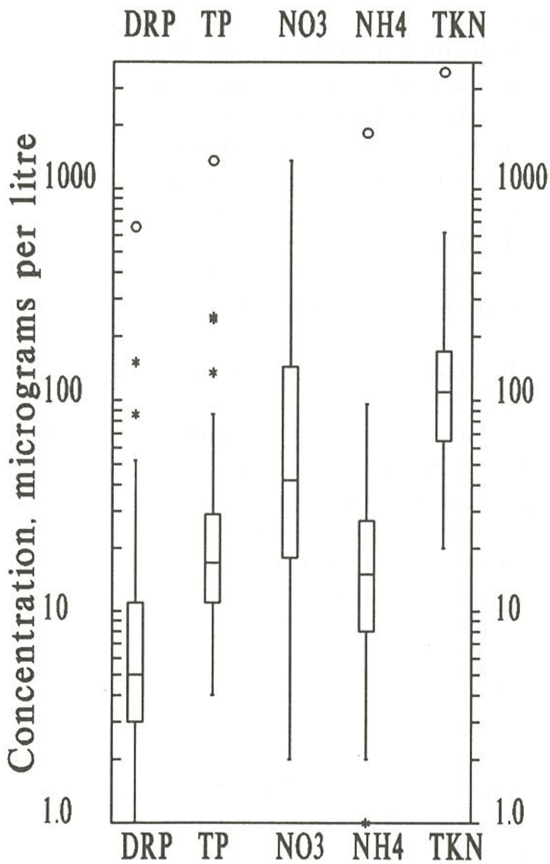

Figure 12.3 Nutrient concentrations in waters of the 100 Rivers study.

Nitrogen. Many of the complexities of analysis for phosphorus are true in analogous ways for nitrogen. However, it is even worse for nitrogen, because nitrogen can exist in water as the inorganic ions nitrate (NO_3), nitrite (NO_2) or ammonium (NH_4), as well as innumerable organic forms. Several forms of nitrogen in water are listed in table 12.2. The most common form is nitrate, at concentrations of 10 to 5000 mgN.m^{-3}. Nitrite at a few mg.m^{-3} can indicate sewage pollution. In Figure 12.3, the median concentration of nitrate is 40 mg.m^{-3}, but you can see the range is extremely wide, even in apparently unpolluted rivers.

N form	Analysed by:
Nitrate (NO_3)	Reduce to NO_2. Analyse for Nitrite.
Nitrite (NO_2)	Analyse NO_2 colourimetrically
Ammonium	Analyse colourimetrically
Kjeldahl N	Digest sample , to get ammonium-N from organic N. Analyse ammonium.
Total N	TKN + nitrate + nitrite
Dissolved Kjeldahl N	TKN on a filtered sample
TIN= Total Inorganic Nitrogen	NO_3 + NO_2 + NH_4

Table 12.2 Forms of Nitrogen in water

Ammonium can occur in swamps and other low oxygen environments, at concentrations of a few tens of mg.m^{-3}. In geothermal springs it can reach 10,000 mg.m^{-3}. In Figure 12.3 the median ammonium concentration is about 15 mg.m^{-3}.

Note that the total Kjeldahl nitrogen (TKN) includes any ammonia in the original sample, but does not include the nitrate or nitrite. In Figure

12.3 the median TKN concentration is about 100 $mg.m^{-3}$.

As for phosphorus, analytical techniques for nitrogen vary. Catalysts used in the TKN test vary greatly, and can give different results on some kinds of samples. Methods should be recorded where they will remain accessible to someone who finds water quality data in 20 years time.

Organic Substances

There is a huge variety of organic substances that may be detected in water. Many of them are toxic substances, with potential harmful effects at very low concentrations. They may result from pollution by industry or agriculture. Analysing for these, and interpreting their effects, is a very specialist task.

At our present level of technology, it is very seldom that the individual organic compounds are seen to be a matter for analysis. When such analysis is needed, it is usually a result of contamination from a known source.

Levels of trace organic compounds in New Zealand are low compared to some European data. Trace organic compounds, including polycyclic aromatic hydrocarbons and polychlorinated biphenyls, have been measured in rainfall, fog and streamwater samples collected in Westland and Fiordland (Herrmann, 1987). The concentrations were one to two orders of magnitude lower than those found in samples from northeastern Bavaria (Germany). Analysis of sediment cores from Lake Tekapo did not show any change in chlorinated hydrocarbon concentrations over the last 20 years, in Herrmann's study.

Organic compounds are usually important because they are used as food by bacteria in the water. In carrying out their metabolic processes the bacteria consume oxygen from the water. This reduction in oxygen concentration leads to a number of undesirable changes to the aquatic ecosystem, especially if the oxygen concentration falls below about 2 $g.m^{-3}$.

The consumption of dissolved oxygen by bacteria in water is used as a technique to assess the amount of organic matter in water. Contaminated water is mixed with uncontaminated water of known oxygen content, and allowed to stand, usually for 5 days. The dissolved oxygen content is measured after this period, and the change in concentration is used to estimate the Biochemical Oxygen Demand of the water (BOD or BOD5). Note that it incorporates any chemical oxygen demand, from sulphides or other chemically active substances. The BOD5 is the decrease in oxygen content of water that would occur over 5 days in the absence of any outside source of oxygen.

The BOD5 may not give a picture of the total oxygen demand of the water, since some wastes take much longer than 5 days to break down. It is easy to get results like <1 $g.m^{-3}$ or >100 $g.m^{-3}$, because of insensitivity of the method, and wrong guesses as to the strength of the original waste.

In uncontaminated waters BOD5 is always much less than 1 $g.m^{-3}$. Values as high as 5 $g.m^{-3}$ can give reason for concern. Some wastes (such as milk) can have BOD5 concentrations over 100,000 $g.m^{-3}$, so a relatively small volume of waste can affect even a large river like the Waikato. Waters which continuously contain large concentrations of some kinds of BOD, such as sugars, develop populations of filamentous brown bacteria known as "sewage fungus".

The BOD is useful, because it measures the likely effect of a waste on the environment. However, it takes 5 days to get an answer. A test known as the Chemical Oxygen Demand (COD) can give a quick result, which for a given waste can be correlated with the BOD. Different wastes or contaminated waters will have different relationships between BOD and COD.

Metals

The term metals, or heavy metals, is normally used to refer to elements like mercury, cadmium, chromium, nickel, arsenic, copper, zinc and lead that are toxic to animals in quite small concentrations. Smith (1986) has reviewed the presence of these substances in New Zealand waters. They are not normally present to any harmful extent in water in NZ, except that geothermal waters can contain a significant concentration of mercury and arsenic

(Smith, 1985) (Table 12.3). Metals may be present in natural streams with highly mineralised waters such as those of the Coromandel Peninsula, or in waste from mining operations, but the practical significance of this is easy to overestimate.

As	Cd	Cu	Cr	Hg	Pb	Ni	Zn
5000	0.5	370	3	2.1	8	23	2

Table 12.3 Maximum concentrations of metals in New Zealand geothermal waters, in mg/m^{-3} (Smith, 1985)

As with other aspects of water chemistry, it is vital when interpreting the information to be clear about the chemical form of the substance to be analyzed. A sample may be filtered to allow an analysis of the dissolved fraction. It may also be digested to ensure all of the metal in the sample is measured.

A measure being promoted by the US Environmental Protection Agency (USEPA 1986) is to carry out a mild acid digestion of a whole (unfiltered) sample. This is intended to represent the "available" fraction of the metal. It is this fraction of the sample that is used when analysing the concentration of metals during toxicity tests on representative organisms, so that both the standard concentration limits and the field measurements can refer to the same property of the water.

It is especially difficult to interpret measurements of the heavy metals. Most compounds of these metals are highly insoluble in water, so the metals that enter a stream are likely to be absorbed on to particles, many of which will be trapped in the stream or lake bed. Measuring a water sample may give a poor indication of the effect of a discharge of heavy metals on organisms, since much of the metal that is ingested by organisms is taken up along with silt particles from the lake or stream bed. The metals also often accumulate in organisms, so an animal may accumulate a significant dose, even from occasional exposures to intermittent sources, when most of the time the concentration is very low.

In the case of mercury, organic forms are far more toxic than the metallic forms, and can be formed by biological action on mercury in the sediments. Therefore, it may be much more important to control the total load of mercury to a stream than to control the instantaneous concentration in the stream.

Other Elements

Many elements can cause potential harm in water. For example, boron can restrict the growth of sensitive aquatic plant species, and is undesirable when irrigating some fruits. Sulphide wastes from some industries are highly undesirable because of both direct toxicity and their effect on oxygen content. Dischargers and water managers need to be alert for possible but unusual situations.

Microbiological Impurities

Water can become contaminated with a large number of microbiological impurities. The most commonly recognised ones are the various kinds of bacteria, such as coliforms (or specifically, faecal coliforms), enterococci, or streptococci. These bacteria can cause intestinal illness, but they are more important as organisms that are relatively common and safe to grow in culture so they can be detected, while indicating that other, more dangerous organisms **may** be present. In the last few years more attention is being paid to viruses, and also to parasites such as Giardia. They certainly affect the use of water, but it is impractical within the scope of this chapter to deal properly with this topic. Interested readers should refer to McBride et al. (1992).

Physical Properties

Colour. People like their water to be blue in colour. Often the presence of organic matter makes the water a brown colour, as is especially noticeable on the South Island West Coast, or in water draining from Waikato peat swamps. This water is not necessarily "polluted" – it may be coloured by natural processes. Davies-Colley (1991) has recommended that to meet the requirements of the Resource Management Act (1991), a "conspicuous" change in colour, caused by a discharge to water, should be defined as a change of not more than 10 points on the Munsell scale, away from that occurring naturally.

Colour can be measured (usually on a filtered sample to get rid of turbidity) by measuring the optical absorbance at a particular frequency in a laboratory spectrophotometer. However, there is little standardisation on just which frequency to use. Timperley (1985) advocated a frequency of 270 nm, in order to maximise sensitivity of measurement, but Kirk (1976) recommended the use of 440 nm in order to correspond to that wavelength which is most strongly absorbed by chlorophyll in living cells. Measurements at different frequency cannot be directly compared.

Visual clarity The traditional measure of clarity of water is the Secchi Disk, a black and white disk that is immersed vertically to a depth at which it just disappears from view. Clear lake water can have Secchi disk depths about 3 to 4 metres. When the Secchi disk depth is less than 1 metre, you are aware that the water is unclear. More recently, Davies-Colley (1988) has advocated the use of a black disk viewed horizontally. This is more in tune with the principles of physics, and can be used in the commonly shallow and clear water of New Zealand streams.

Light penetration in water When water is both coloured and turbid (as on the West Coast of the South Island) the low light levels on the bottom of the stream can have a marked detrimental effect on the flora and fauna, because of the reduction in photosynthesis. (Quinn et al, 1991) The measurement of light penetration is described by Davies-Colley (1991) It is different from clarity, because for clarity, the light must not be scattered. A body of water, like that of Lake Tekapo, may be highly turbid, so that you can see only a short distance in it. It is therefore not very clear. However, the particles in the water may not absorb light very well, so that the amount of light at a depth of several metres may be quite high.

Temperature The Resource Management Act (1991) requires that waters being managed for a number of ecological purposes must not have their stream temperatures changed (by reason of a discharge of water to them) by more than 3 degrees Celsius. This includes decreasing the temperatures of geothermal streams. The main situation where temperature change occurs is in the cooling of power stations, but it has also arisen when considering the discharge of drainage water from a coal mine into a small stream, where the underground water was constant at about 12 degrees all year round. (Unpublished water right hearing, Waikato Valley Authority.)

Increases in stream temperature can have a dramatic effect on freshwater fisheries by decreasing dissolved oxygen levels and causing changes in the stream biota (Brown et al., 1971). As well as being a result of some discharges of water to the stream, very significant temperature increases can result from clearing vegetation that previously overhung the stream. It may be a legislative anomaly that temperature changes as a result of discharges are prohibited, but changes for other reasons are not.

Measurement of Water Quality

What to Measure

We assume that water quality, in the context of a particular use, can be defined in terms of the water physics and chemistry, as described above. There are very many properties of a water body that could be measured in order to define its water quality. In order to define a list for a particular survey, it is necessary first to define the purpose of the survey, and then to select from the large variety of possible measurements a set of them that is feasible to measure, and which is most relevant to the purpose. (Hoare, 1983a,b) For example, in a study of lake eutrophication, measuring phosphorus and nitrogen species would be more important than measuring heavy metals or BOD.

How Precisely to Measure

Many analytical techniques are often available for a given parameter and can be found in standard texts (APHA/AWWA/WPCF, 1989). The choice of technique often depends on the expected concentration of the parameter, the laboratory facilities available, finance and staffing.

Measurements must be precise enough to enable a conclusion to be drawn. If you are worried about heavy metal contamination of the Waikato River, and the concentration of concern for say, chromium, is 10 $mg.m^{-3}$, then it is not very helpful to make all measurements with a detection limit of 20 $mg.m^{-3}$. If you have a bacterial contamination problem at a river beach, and you conduct a survey of the river to find out the relative importance of various sources, don't use a lab that reports anything over 2400 bacteria per 100ml as ">2400". (Such a report is fine to decide that a beach is unsuitable for swimming, but it does not allow you to compare various tributaries that are all contaminated to some extent.)

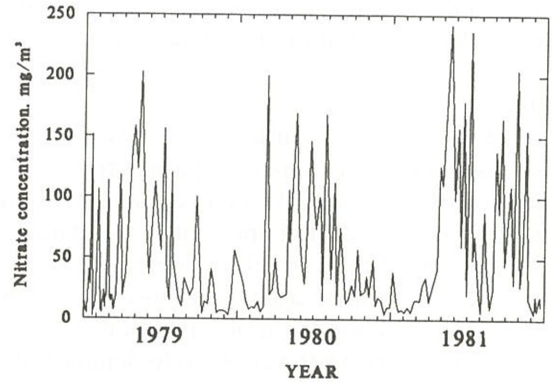

Figure 12.4 Nitrate concentrations in the outflow from Lake Rotorua. (Hoare, unpublished data.)

Statistical Considerations

When trying to define the water quality of a lake, river, groundwater or the sea, you cannot take a single sample and expect to know much about the water quality. In many cases there are strong, apparently random, components in the variation in concentration of constituents. In other cases there are clear diurnal or seasonal variations (Figure 12.4). Frequently, the concentrations of constituents vary with the flow rate of a stream (Figure 12.5).

If the regular (seasonal or diurnal) variations are unimportant in a particular situation, then the more samples you take, the closer the average of the samples is likely to be to the "true" average of the water body. The standard deviation of the mean of a set of samples is equal to the standard deviation of the raw data, divided by the square root of the number of samples. In principle, you can make the mean of your set of samples as close to the "true" value as you like, but in practise the gains become quite small after the first 10 samples or so.

It is not uncommon for values of a series of water quality samples to have frequency distributions that are skewed towards the lower levels, or, in other words, to have occasional very high values. This is why in figures 12.1 and 12.3 logarithmic scales have been used. When it is necessary to compare average concentrations,

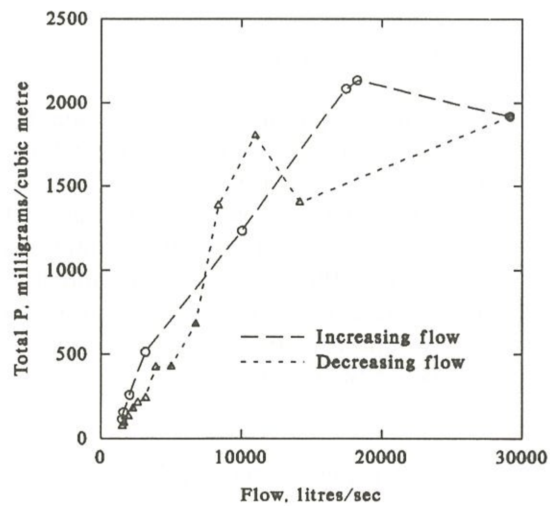

Figure 12.5 Total Phosphorus concentrations in a large flood of the Waiteti Stream, Rotorua. (Hoare, unpublished data)

for instance, when similar catchments have different land uses, there are statistical tests that are often used, but which may not be applicable to this sort of data distribution. (Students t-test, for example, requires the values to be normally distributed.)

One solution to this difficulty is to transform the data, by taking logarithms. Non-parametric, or distribution free, statistical techniques are also useful and may be especially important if the data set contains data below or above detection limits. However, specialist advice should be sought before venturing too far in this direction.

Nonparametric tests were used to demonstrate, for the Purukohukohu experimental basin in the central North Island, changes with time in the concentration of streamwater phosphorus and nitrogen species in forested and pasture catchments (Cooper et al., 1987).

Rutherford (1983) showed how to design a survey for maximum efficiency when studying lake water quality, including the possibility of incorporating a model to allow for seasonal variation. When designing large surveys the incorporation of statistical techniques can save a lot of money, or greatly improve the usefulness of the data. (Rutherford 1984.)

Timing of Samples - Seasonality

Many water quality parameters vary seasonally. (For example, nitrate in the outflow from Lake Rotorua - Figure 12.4.) This means that surveys of these parameters must be conducted over at least one year. A minimum of six 2-monthly samples is required to define the seasonal cycle, and it is much safer to have 12 monthly samples to catch short-term variations, and to allow for some unusable samples in the period.

Relationship Between Water Quality and Flow Hydrology

The concentration of contaminants may vary with flow rate, so that it is important to cover the full range of flows. Dissolved constituents can either increase or decrease during floods (Claridge, 1970, Hoare, 1987). Often there is hysteresis, where concentrations are higher at a given flow rate on the increasing flow than they are at the same flow when flow is decreasing. Figure 12.5 is an example of the variation in total phosphorus concentrations during a flood (Hoare, 1982). When both the concentration and the flow rate increase like this over several orders of magnitude during a flood, the amount of material transported can be very large. In this situation one cannot estimate the effect of floods on water quality without carrying out detailed studies of the concentration during floods.

Frequently, it is necessary to estimate masses of material carried by streams, and to do this you need to combine the concentration data with flow information, by multiplying these two measurements and integrating over the flood duration. For example, to estimate the importance of leachate from a land disposal site on a receiving water, you need to know the **amount** of leachate. It can often be convenient to install a small measuring weir at the place where water samples are collected, and to collect flow information at the same time as collecting the samples.

Synoptic Sampling

This kind of sampling occurs when you wish to understand the processes occurring in a stream system. You need to take samples from a set of locations that in some way are connected. For example, when you want to know which are the major sources of bacteria or BOD in a stream with multiple sources and tributaries, you would try to take samples from as many places as possible in a short time, and you would then identify the places making largest contributions. You usually need extensive information on stream flow to enable them to be fully interpreted.

Often these surveys are conducted to gather data for a modelling exercise. One of the best New Zealand examples is a survey of the Tarawera River, to gather data used to calibrate a dissolved oxygen model. (Rutherford et al. 1991)

Natural Sources of Major Impurities

The atmosphere provides dissolved oxygen (10 $g.m^{-3}$) and bicarbonate ions (40 $g.m^{-3}$). The next greatest sources of impurities are the rock and soil that the water passes through on its way from the clouds to a river or lake. This is where ions such as sodium and calcium, and silica, come from. Some major ions also come from the sea, via dust particles and aerosols. These loadings can be significant near the coast. Blakemore (1953) measured chloride inputs (as NaCl) of nearly 400 kg/ha/year at Rongatai, but the input dropped to 60 kg/ha/year at Te Awa, 55 km from the coast. Sulphate and magnesium inputs can also be significant, which reflects the sea-water ionic composition.

There are natural sources of heavy metals and other undesirable substances. (Smith, 1985). The geothermal springs found mostly in the Central North Island, but also in Northland, and along major fault zones, contribute locally significant concentrations of heavy metals and sulphide (Table 12.3). Fish in the upper Waikato River contain mercury at an undesirably high level because of contamination from local geothermal springs. (Brooks et al, 1976).

The swamps found all over New Zealand have water that is very high in organic compounds. Dissolved organic carbon concentration (including fulvic and humic acids) can be as high as 37 $g.m^{-3}$ (Collier et al., 1989). Left to themselves swamps do not cause problems - they have their own specialised ecosystems. However, when they are drained the highly coloured, low oxygen content water leaving them may have a very large effect on the water body receiving the drainage water.

Diffuse Man-made Sources of Impurities

We often think of the impurities attributable to people as being incorporated in a water discharge from a pipe. However, in NZ the biggest effects of people are those associated with their changing the land use from the native forest to other uses. These are **diffuse** effects, and may be seen as increased concentrations of nutrients in the water, decreased oxygen content, high bacteria content, or raised temperatures.

Forestry

The main effect of forestry is to allow solids to enter the streams during certain phases of the production cycle. A study of forest erosion in Nelson (Fahey and Coker, 1989) indicated that sediment production rates were about 320 tonnes/km^2 when road construction rates were near their peak, but that they subsequently decreased to 10% of this figure. The major sources of solids in this erodible granite terrain were cut banks, drainage ditches and side cast fill near gullies and streams. Road surfaces were a minor source of solids.

An experimental catchment at Maimai illustrates some of the water quality problems associated with forest operations. (Table 12.4, from data of O'Loughlin, et al., 1980.) Despite the riparian reserve with live vegetation acting as a filter, storm runoff from the tracks in M9 carried concentrated sediment loads which were diverted in channels to the stream. Forestry operations can also result in the loss of dissolved substances, particularly in the first year or two after burning (Rowe and Fahey, 1991).

Catchment Number	M7	M9
Logging Method	Log hauler	Log skidder
Burning?	Yes	Yes
Riparian Reserve?	No	Yes
Solids yield	1.4 times control	8 times control
Solids type	Ash	Erosion from track and sidecasts.

Table 12.4 Results of experimental logging

Careful planning of roading and the use of appropriate logging techniques can make erosion a relatively minor water quality problem, especially if the effects of logging over a long period are compared with the effects of other forms of commercial land use over the same period.

Grazing

On flat land, grazing has little effect on erosion rates. However, animal wastes can affect surface streams if the streams are not fenced, allowing stock to walk in and defecate in or near them. Effects include an increase in suspended solids, bacterial counts and BOD.

Grazing animals deposit urine in concentrated spots, and it has been shown that microbiological processes in the soil convert the organic nitrogen in the urine to nitrate, which can be flushed into the groundwater by rain or irrigation water. (Burden, 1982.) This process can yield 30-70 kg/ha/yr of nitrate-N to the groundwater, and lead to concentrations of nitrate-N up to 30 $g.m^{-3}$. Since the WHO limit for nitrate-N in drinking water is 10 $g.m^{-3}$, this is a matter for concern. In practice, there is little that can be done to prevent the problem, and the main corrective mechanism is to avoid using such water to make up formula for infants, the people most likely to be affected. The main areas affected are the Waikato and Canterbury plains, but there are many other local areas of concern.

On hilly country, groundwater problems are usually not significant because of lower stock carrying capacity. However, losses of both nitrogen and phosphorus from hill country to surface streams can be significant, especially if the land has been fertilized. The majority of the losses occur during storm events, and pasture catchments tend to have larger losses than nearby forested catchments (McColl, et al., 1977; Cooper and Thomsen, 1988).

Retiring riparian (streamside) zones from grazing can substantially reduce (by 55 to 87%) the amount of nitrogen and phosphorus species and solids exported into streams. This is attributed to the filtering action of a dense vegetation cover (Smith, 1989). However, Cooper et al. (1990) found that diffuse nitrogen and phosphorus loads on Lake Rotorua would be reduced by only about 9 and 17% when the Upper Kaituna soil conservation measures are complete.

Horticulture

This is usually carried out on flat land, where erosion is no problem. However, in some places, such as at Pukekohe, very fertile volcanic soils occur on hillsides, and where they are farmed for onions or potatoes there can be locally severe erosion when intense rain occurs after ploughing.

Some popular concern has been expressed about the possibility of agricultural chemicals being lost from horticultural operations to surface or groundwater. This is very difficult and expensive to investigate, but Close (1992) has made a start. Out of 82 wells selected from seventeen areas around New Zealand as being in a place most likely to be contaminated, pesticides were detected in 9. One sample in each of 2 wells exceeded the most conservative international guideline, while the rest were very low. The results show that it is possible for groundwater in unconfined aquifers to become significantly contaminated by pesticides.

Urban

The main potential water quality problem arising from cities is sewage (a point source problem). In addition to this, there are instances where stormwater runoff is seen to be a problem. Several of the regional councils have noted this in a survey conducted in conjunction with the preparation of this chapter, the results of which are discussed later.

Stormwater from urban areas has been described in overseas literature as being similar in quality to sewage. (Whipple et al, 1974, Cordery, 1977). If you compare untreated stormwater from an urban area with the secondary sewage from the area, then the peak stormwater concentrations can appear to be worse than those of the sewage. However, if you compare the total loads of phosphorus with those from rural (pasture) catchments, the land uses have similar effects. (Hoare, 1984). It is likely that the load of other constituents would also be similar, for pasture and urban land uses. If a river, lake or estuary receives water from both urban and rural runoff, little improvement in its water quality may be found from treating just the urban runoff contribution.

A factor in urban runoff is the use of leaded fuels, which increases lead concentrations in the streams. This problem is being tackled by a change in fuel types. Other potential problems arise from zinc in vehicle tyres.

Acid Rain

In some parts of the world large quantities of sulphur are emitted by smelters and by power stations. The sulphur oxides combine with rain to form various sulphur-based acids. Combustion engines in vehicles emit nitrogen oxides, which form nitrogen-based acids when dissolved in water. Although these processes occur in NZ, the nature of our weather systems, and our small size, mean that noticeable adverse effects are not caused to our waters. Instead, the pH of our rain is about 5.6, due to dissolved carbon dioxide (Verhoeven et al., 1987).

Point Sources Of Impurities

Sewage Waste Treatment Systems

These are the most obvious sources of potential water quality problems, such as bacteriological contamination or excess plant nutrients (eutrophication), but they are not necessarily the most important. It is always possible to treat sewage to reduce potential nuisance chemicals and organisms. Problems arise in deciding to what level the treatment should be taken, and finding the money to meet the desired level.

Philosophical arguments abound. Should we be able to use the receiving water body to treat the waste to some extent? Should we always dispose of waste through a land filter? Should we use up valuable land resources, minerals and energy for a treatment plant, or should we let the sea process the waste naturally, becoming fertile so that more fish will grow?

The Resource Management Act, 1991, makes it clear that our water resources should not be prevented from being used for many purposes because of the presence of poorly treated sewage wastes. A survey of Regional Councils (undertaken for this chapter and reported later) showed that in general, they believe that sewage wastes are under control.

Table 12.5 shows the methods of treatment in use in the 30 largest cities in New Zealand, in 1991. (Cocks and Patrick, 1991) The substances in the waste that affect water quality depend on the treatment system, but BOD and suspended solids are always of concern, and sometimes pathogenic organisms are likely to be present. High concentrations of ammonia in the waste can affect fish, even after great dilution. Local bodies usually take care that industry does not put toxic substances into sewers. Smith (1985) concluded that heavy metals are unlikely to be an important component of municipal sewage wastes.

Treatment type	No.
None (Ocean outfall)	4
Preliminary Treatment	4
Primary	4
Secondary	15
Tertiary	6

Table 12.5 Types of sewage treatment in use in 1991

Landfills

Solid wastes can be a source of surface or ground water pollution, if landfills are not properly managed. Willmot (1991) has reviewed the concentrations of substances in landfill leachates. The effect of the waste in any given situation will depend on the dilution and adsorbtion of any waste that escapes from the landfill.

Industrial Wastes

The largest sources of industrial wastes in NZ are the plants processing our agricultural products. Many of the potential problems are similar to those of sewage treatment plants. Feasible ways to treat this waste are almost always available, but they are not always used, partly because factories discharging waste before 1963 have been allowed to continue their unsatisfactory practices. The Resource Management Act (1991) has now given Regional Councils more powers to enforce modern waste treatment methods.

Fellmongeries and tanneries use industrial chemicals with special treatment problems (e.g., sulphides, and chromium). The timber treatment industry has used pentachlorophenol (PCP) for many years, but it is now recognised that this chemical may contain dioxin as an impurity, and that casual waste collection and treatment practises may have led to potentially serious environmental contamination. A host of small industries, including the pesticide and other agricultural chemical industries, have a potential to contaminate water with a wide range of substances. Predicting what should be looked for in any individual situation requires knowledge of what might be released, since it is impractical to search for every chemical in industrial use.

Rural Wastes

Point source rural wastes are mostly from dairy sheds and piggeries. Treatment for BOD load is well researched, but in many cases the maintenance of a treatment system, even when it is installed, is poor or non-existent. The Waikato Regional Council alone has identified over 4700 rural point discharges (many of which are intended to be to land, but which actually reach streams). Of all point sources, rural waste has by far the biggest adverse effect on the water quality of NZ's water (Hickey & Rutherford, 1986).

The effects of rural point source discharges are normally controlled by limiting the BOD load, thereby maintaining oxygen in the streams. However, even when this is done properly, ammonia concentration in the receiving water can be high enough to cause fish toxicity, or a barrier to fish migration. (Hickey *et al.* 1989)

Mining Wastes

Much public discussion is generated over the waste generated by the mining industry. Overseas, there are very impressive examples of lunar landscapes created by mines. In NZ, we have coal mines and metal mines, both of which have led to pollution problems in the past.

The open cast coal mines at Huntly created huge but unmeasured loads of silt to Lake Waahi and the Waikato river in the 1970's. Some of the water right procedures designed to control these problems led to very long and expensive tribunal hearings. During the 1980's there were changes in attitude of many of the parties concerned, and now it seems that each mine has a properly designed and maintained treatment system, and the water right conditions are treated seriously. (Waikato Regional Council staff personal communication.)

On the Coromandel Peninsula there is a lot of opposition to gold mining, on philosophical and social grounds as much as anything else. It is often claimed by those opposing such mines that the waste cannot be treated safely. Large amounts of very fine rock fragments are created, often highly mineralised, and therefore potentially toxic. These must be stored for a very long time in a high rainfall area, so some sophisticated techniques must be used. Leachate from the tailings dump may escape.

However, for the same reason that the mines are there, the area has naturally high concentrations of heavy metals in some of the streams (Livingston, 1987). Barring disasters, the stream water quality after modern mine waste treatment is unlikely to be much different in the presence of the mines than without them. However, the possibility of such disasters should be not be taken lightly. We have had two hydroelectric schemes fail disastrously in the central North Island in the last decade, with locally catastrophic effect on the aquatic environment.

Heat Sources

Thermal power stations must dispose of about half of their total energy consumption to the environment, because of thermodynamic considerations. This is often done by using rivers as heat sinks. In order to meet the legal requirement that temperatures be increased by less than 3 degrees, the output from the station may be limited at low river flow. Also, it is likely that a maximum water temperature will be set, as at the Huntly power station (25.9 degrees Celsius), and this may limit power output in summer.

Because power stations are large, predictable machines, they are unlikely to be concealed, and their effects on temperature are (relatively) easily predictable. Predicting biological consequences of a temperature rise is not so easy, though. The setting of limits on maximum temperature can involve detailed behavioural studies on fish or other animals.

Estimating Loads of Contaminants.

General

In order to manage water resources it is frequently necessary to evaluate the loads of the many substances that find their way into our water. For instance, to decide on the importance of sewage loads to Lake Rotorua, all the other sources of plant nutrients had to be evaluated. Since the problem was deemed to be important, and our level of knowledge was low, a major exercise was undertaken to measure directly as much of the total load as possible (Hoare, 1987). With the knowledge that resulted from 3 years of intensive study, it was possible to compare the loads from sewage with those from all other sources, and to make predictions about the water quality resulting from reducing the sewage load to various levels (Hoare, 1980).

However, it is usually not necessary to go to this extreme of effort to obtain enough data to make management decisions on the importance of waste loads. There are various compilations of expected yields per unit area, concentrations of substances in waste, nutrient or BOD loads per capita and so on, as mentioned below. These can be used to solve many problems just as a desk exercise. Even when it is expected that measurements will be made, a desk exercise can be very useful in planning a sampling strategy.

Diffuse Sources.

The yield of potential water pollutants from land surfaces is found to be very dependent on the use of the land (Wilcock, 1986). In many cases one can use a compilation of measurements such as that found in Wilcock (1986) to obtain yields of nitrogen, phosphorus or suspended solids per hectare. Some results are shown in Fig. 12.6. The possible range of such values is wide, but by going back to the original sources of data, (e.g., Cooper & Thomsen, 1988) a situation relevant to a particular case may be found. Don't expect a measurement to be more precise than to be within a factor of two of the right answer. (It is difficult to do better than this even by measuring loads directly!)

The difficulties with making measurements of diffuse source yields include coping with extremely large flow-related variations in concentration. If a source of pollutants enters a lake, the lake can act as an "averaging" device, which smooths out the

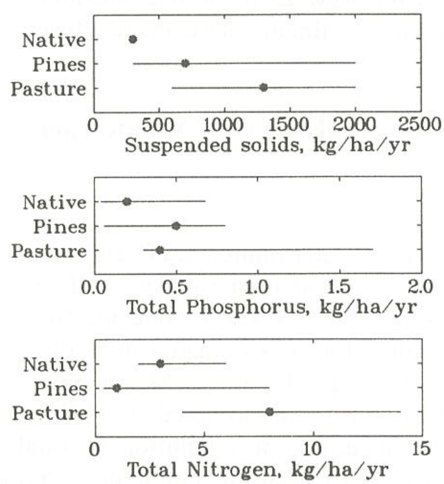

Figure 12.6 Range and median of the yield of solids, phosphorus, and nitrogen from pasture, active pine forestry, and native forest. (Data from Wilcock, 1986.)

variations in concentration, and removes the correlation with flow rate. Relatively few simple measurements can give the amount of a substance leaving a lake each year. Of course, allowance must then be made for the possible retention of substances in the lake.

Vant & Hoare (1987) show how it is possible to use measurements on lake water quality and the hydrology of a lake to estimate the total load of phosphorus on the lake. In most cases in NZ this will be a diffuse load.

Point Sources

A common procedure for estimating point source sewage loads is to use figures on per capita nutrient or BOD loads to estimate the input to a sewage treatment plant, and to assume that a given level of treatment is reached. This treatment can be expressed as percentage removed (nutrients), or as expected outflow concentration (BOD, bacteria or suspended solids). For example, Hoare (1984) estimated that in Rotorua, N and P loads on septic tanks were about 6 and 1 kg per person per year, respectively, in the early 1980s.

Similar techniques apply to animal wastes. Cooper et al (1979) gives a useful range of concentrations for New Zealand slaughterhouse effluents, and some estimates of treatment efficiencies.

Estimating the Effects of Waste Loads

Dilution

Most criteria for environmental effects are based on the concentration of a substance. If greater dilution is available, a lower concentration will result, and therefore more waste can be disposed of. According to the Resource Management Act (1991), dilution is to be estimated after allowing a reasonable "mixing zone". Dilution is usually a straightforward calculation, but it does require that you have some hydrological knowledge to start with. Collecting this at the same time you collect the concentration information is sometimes not done as well as desired.

Note that floatable material will not usually be subject to dilution effects.

Self Purification

Natural processes that result in improvements in water quality with time (in the absence of continuing pollution) exist in all waters. Sunlight can sterilise the water. With free access to the air, bacteria can oxidise most wastes to harmless substances. Clays and soil particles can adsorb heavy metals. Bacteria can denitrify nitrogen compounds to nitrogen gas.

Some groundwater disposal systems treat the waste effectively while the water percolates through the soil, but in many cases some constituents, such as nitrate (Burden 1982), enter the groundwater and persist for a long time, potentially affecting other uses. If, after years of discharge, a component that affects the use of the groundwater is found to be present in the waste, then we will be left with a groundwater that may take hundreds of years to clean itself up. NZ rivers will flush themselves clean in weeks.

Many of NZ's groundwaters are very high in nitrate. However very few of the rivers are, despite being fed from these groundwaters. It has been shown that both bacterial denitrification, and uptake by plants in the river are responsible for this (Howard-Williams et al, 1986).

The decay of the oxygen-consuming substances in streams has been well researched over the years, and is still being studied. Very good predictions of the dissolved oxygen remaining in a river subject to organic discharges can be made (McBride & Rutherford, 1984). NZ streams have proven to have rather more rapid self-cleansing properties than many of those studied overseas, so models that have been found to work in NZ should be used in any work that is carried out here.

Toxic Levels

When deciding what level of any substance is acceptable in the environment, one can refer to a number of compilations, such as the series of publications put out by the USEPA. Table 12.6 summarises some of the standards from USEPA (1986). It is important for users of this kind of information to read the rationale behind the numbers, and to keep up with the successive revisions of these criteria.

Chemical Species	Toxicity (freshwater)		Human Toxicity
	Chronic	Acute	
Arsenic III	190	360	0
Arsenic V	U	U	0
Cadmium	0.66†	1.8†	10#
Chlorine (Residual)	11	19	-
Chromium VI	11	16	50#
Chromium III	120†	980†	-
Copper	6.5†	9.2†	1000
Cyanide	5.2	22	200#
Iron	1000	1000	300#
Lead	1.3†	34†	50#
Manganese	-	-	50#
Mercury	0.012	2.4	0.144
Nickel	88†	790†	-
Nitrate	-	-	10#
Silver	0.12†	1.2	50#
Sulphide (H_2S)	2	-	-
Zinc	59†	65†	-

Notes:
† At a hardness of 50 g.m^{-3}. Lower toxicity at greater hardness.
U Unknown, but more toxic than As III.
US Drinking water standard.

Table 12.6 Toxicity standards, mg/m^3, taken from USEPA

The USEPA allows for the fact that organisms may be able to tolerate high concentrations for a short while, say an hour or so, (the acute limit), but for chronic exposure, through a whole life cycle, a lower limit is needed. Human drinking water standards are generally less stringent than the concentrations needed to protect organisms which spend their lives in the water.

The USEPA places a strong emphasis on the need to consider the toxicity to particular organisms likely to be present in the water under consideration. While they give general figures, at the same time they say that users of the guidelines should test their own organisms for their tolerance to various chemicals.

Those setting limits on concentration must take account of the differences between acute and chronic toxicities, the statistical nature of concentration variations, and the practicality of sampling to determine the existence of an undesirably high concentration. They must have a precise understanding of the effects of different chemical forms of the substances, and the effects of different pH or water hardness. A list such as Table 12.6 can be used as a guide to the likely importance of a measurement or estimate of a concentration of a toxic substance, but in a marginal situation professional advice must be sought.

Microbiology

The NZ Department of Health has recently released guidelines for microbiological organisms in water. (McBride et al, 1992.) The approach taken to setting standards follows that of USEPA (1986). Upper limits are set for the median concentration of 2 organisms over the whole of a bathing season, and for the maximum value of the concentration of organisms in a single sample. The medians over the bathing season should be no more than 33 enterococci per 100 ml, or 126 *E. coli* per 100 ml, in fresh water. For single sample concentrations, the value depends on the intensity of use of the water, as shown in Table 12.7.

A Water Quality Index.

It should be clear by now that determining water quality is a complex business. It is easy to measure many aspects of water quality, but how can something be synthesised that helps people quantify their "gut" feeling?

A number of organisations have developed water quality indexes, that combine by numerical

Upper limit per 100 ml	Designated bathing area	Moderate use	Light use	Infrequent use
enterococci	61	77	107	151
E. coli	235	293	410	576

Table 12.7 Maximum concentrations for a single sample

methods the values of a set of key parameters, and come up with a single number. Certain ranges of this number indicate suitability for a given use, and other ranges indicate potential problems. Each use may need a different set of parameters, or use different "weighting" factors.

Smith (1990) has developed a set of indexes for New Zealand waters, using a panel of experts to ensure that it reflects New Zealand practical experience. The water uses covered include "General", "Regular Public Bathing", "Water Supply", and "Fish Spawning (salmonids)". They are based on measurements of dissolved oxygen, pH, suspended solids, turbidity, temperature, faecal coliforms, BOD, and ammonia. The index values enable you to answer such questions as "How suitable is a particular water for certain uses?". They are also useful as a way to compare locations, or to keep track of short-term changes.

Table 12.8 (from Smith 1990) is an example of how the index can be applied to a set of water quality observations. The observations are turned into index values for the "General" water use class by means of relationships graphed in the publication, and the overall index score is taken as the **lowest** sub-index value, in this case 36. A score less than 40 indicates that the water is not suitable for this use.

Since the original work, Smith and Davies-Colley (1992) have extended the method to include assessments of whether the visual clarity is suitable for bathing or aesthetic purposes.

Lake Modelling

Much of the early literature on this subject was built around attempts to classify subjective assessments of American and European lake water quality, and the early OECD studies (Vollenweider, 1968) in particular failed to take full advantage of the easily obtained information on lake nutrient concentrations.

A modern approach to water quality problems of the type encountered in NZ is found in the "Lake Managers Handbook" (Vant, 1987). This is based on an empirical relationship between the self cleansing effect (the phosphorus retention coefficient, R) and the lake hydrology. The relevant equations are

$$P_{lake} = P_{input}(1\text{-}R)$$

$$R = 15/(18 + Q/A)$$

where P_{lake} is the average phosphorus concentration in a lake, P_{input} is the annual average phosphorus concentration in the inputs to the lake, R is called the retention coefficient, and is defined by the relationship between P_{lake} and P_{input}, and the second equation is an empirical (and not very precise) relationship between R, the annual

Measurement	Value	Sub-index value
D.O.	7.7	87
pH	8.5	83
Temp.	23.5	73
SS	24	51
Turbidity	12	56
BOD	3.5	64
Faecal Coli.	4000	36
Overall Index Value		**36**

Table 12.8 A water quality index calculation

volume of water leaving the lake (Q), and the lake area (A).

When estimating the effects of a waste load on a lake, you usually know the size of the load, but in order to estimate its effect, you need to compare it with other loads, and to estimate the "self-cleansing" ability of the lake itself. From P_{lake} (an easy thing to measure) you can estimate P_{input} (difficult to measure) and hence the total load (Vant & Hoare, 1987). The technique can also be used the other way around, to estimate the lake phosphorus concentration after changing the nutrient load (Williamson & Hoare, 1987).

This modelling work shows that when disposing of a phosphorus load to lakes there are two distinct controlling processes. If a lake has a high hydraulic load (in other words, a large volume of water flows through it, compared to its area) then dilution of the phosphorus will occur (P_{input} is low), and little effect will be noticed. On the other hand, if the hydraulic load is low, then self-purification is very important (R is nearly 1), and up to 90% of the P load can be safely locked up in the sediments, and not affect the lake water quality. This latter effect is unexpected, and is frequently ignored, even by limnologists.

Survey of Regional Councils

In order to ensure that major issues in water quality were not being ignored by the authors, an informal survey of Regional Council water managers was carried out for this chapter. Of the 14 organisations contacted, 13 returned the simple one-page questionnaire. For space reasons, only the questions, and a very brief assessment of the replies can be given here.

What are the good things about your region's water quality?

All regions reported that they had good water in the headwaters of their catchments, often in large quantity. Several had good groundwater.

What water quality problems do you have?

The most common problem (11/13) was one or other aspect of non-point-source pollution, often associated with grazing animals. Sediment and bacteria were explicitly mentioned.

The next most common problem (7/13) was point source pollution in the lower reaches. These were more often accidental or uncontrolled, rather than inadequately controlled.

Four regions reported concern with groundwater nitrate concentrations, and three were concerned about urban stormwater quality.

What do you see as being ways to protect the good water?

Eleven expected to use regulation in one form or another (classification, or planning controls). Eight expected to use land use controls, and five focused on education.

What ways do you intend to use to fix the problems you have?

Education was seen as important by 8 of the councils, followed by enforcement by 6. Other ways were monitoring (4), regulation (3), land use controls (3), and research (1).

Are there public perceptions that need to be changed?

Nearly all councils felt that the public was not sufficiently aware of the problems caused by waste entering water. This was particularly true for urban stormwater and rural waste. Several councils were actively working on this.

Any other comments?

A variety of comments mostly repeated the points made elsewhere. One council suggested research was needed, a point also made by another one under the heading of fixing present problems.

Overall analysis of the questionnaire.

It is clear that the managers of water feel that we still have a lot of very good water in this country. The present management emphasis is on prevention and correction of non-point- source pollution, in both urban and rural situations, presumably because the tools to correct piped discharges are at hand and just need to continue to be used.

Public educational programmes feature as ways to fix present problems, and to protect the good water, since it is felt that the public have yet to learn the importance of keeping water clean.

There is a strong interest in monitoring water quality, as a tool to both fix and protect water quality, but little interest in research.

Acknowledgements

The authors wish to thank Noel Burns and Murray Close for their helpful reviews, and Carmen Kirkwood for assistance with the Maori perspective. Many staff of the Water Quality Centre helped us by providing references and data.

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13
Sediment Load

D Murray Hicks and George A Griffiths

Muddy Rivers - Who Cares?

Within an average year, New Zealand's rivers carry a staggering 400 million tonnes of sediment from the land to the ocean. One particularly dirty river, East Cape's Waiapu, alone carries 35 million tonnes per year - equivalent to a 5-tonne truck-load every 12 seconds. With other rivers of the area, it is responsible for the perpetually muddy coastal waters found off East Cape. The sediment moved during storms and floods can create many problems, ranging from soil erosion and sediment deposition in river channels and reservoirs to diminished water quality through the whole catchment-ocean system.

Sediment problems are of concern to soil conservators, engineers, geologists, planners, and natural resource managers. Typically, these practitioners require information on the sediment yield over a period of time. They might, for example, wish to: predict deposition in hydro-electric power reservoirs; know the supply of sand to coastal littoral cells; identify the main sediment sources in catchments suffering sediment problems; evaluate the effectiveness of soil conservation measures; match regional-scale denudation rates with tectonic uplift rates; measure the flux of pollutant chemical species that "piggy-back" on sediment grains; or predict long-term changes in sediment yields associated with changing climate and land use. Good sediment information is particularly important for the planners of hydro-electric reservoirs, given New Zealand's high reliance on hydro-electric power. While most New Zealand hydro-electric reservoirs have been constructed downstream of natural lakes, which serve as sediment traps, sediment supplies from tributaries can still cause problems that may include loss of operating storage, degradation of the river bed downstream of dams, aggradation with increased flood levels upstream of the reservoirs, and abrasion of turbines and other plant (Jowett, 1984). The annual cost of regaining lost storage can be high, especially when added to lost generating capacity.

Those concerned with environmental issues are interested also in the "instantaneous" sediment load and temporary erosion and deposition. These influence a river's recreational and aesthetic values, the quality of domestic and industrial water, and the impacts of water diversions and mining activities on river-bed stability, water quality, and stream ecology. For example, suspended sediment reduces light penetration and hence primary productivity, thus influencing the composition and abundance of the stream-bed invertebrate community, and in turn the abundance of fish (Ryan, 1991). Bed scour and deposition may also interfere with spawning conditions.

Thus, directly or indirectly, all New Zealanders should care about muddy rivers, and, sooner or later, students and practitioners of the hydrological sciences - be they from engineering, planning, or environmental disciplines - are likely to face a problem requiring information on sediment. This chapter discusses the rudiments of river sediment load: what it is, where it comes from, how it is transported by the flowing water, how it is measured or estimated, its known characteristics and variability among New Zealand rivers, and the natural and cultural factors controlling this variability.

The Sediment Load

The sediment load of a river consists of inorganic material (rock fragments and mineral grains), ranging in size from the finest clays through silt, sand, gravel, and occasionally even boulders, that is entrained by the flowing water. The sediment load is quantified either in terms of its concentration or its discharge. Sediment concentration is the dry mass of sediment per unit volume or mass of water-sediment mixture, expressed commonly in units of milligrams/litre or parts per million by mass, respectively. Sediment discharge is the rate at which sediment is transported past a cross-section, generally expressed in units of kilograms/second or tonnes/day.

Traditionally, there are two ways in which the sediment load has been sub-divided: by source or by mode of transport (Figure 13.1).

By source, the total load is divided into wash load and bed-material load. The wash load consists of particles finer in size than those commonly found in the river bed - typically silts and clays (material finer in diameter than 0.062 mm) - which have been washed into the river from its catchment. Wash-load concentration in the river is largely dependent on its rate of supply and is only indirectly related to the river's flow rate and capacity to carry sediment. The bed-material load, as the name indicates, is derived from the river bed, and is typically sand and/or gravel; its concentration is directly related to the river's sediment transport capacity.

By mode of transport, the sediment load is divided into suspended load and bed load. The suspended load is held in the flowing water by turbulence, moves with practically the same velocity as the water, and is carried for considerable distances without touching the bed. Usually it is fine sediment: in terms of source, it largely is wash load plus the finer fractions of the bed material. The bed load is sediment moving in almost continuous contact with the bed, rolling, sliding, or hopping (saltating) under the driving action of the flowing water. The bed load typically is derived from the coarser fractions of the bed material. Bed load and suspended load may move simultaneously; the borderline between them is not well defined; and sediment that moves as bed load at one flow rate may be transported in suspension at a higher flow rate.

Down a river system, the bed material generally tends to become finer and suspended-load transport increasingly dominates bed-load transport. The two processes responsible for these downstream changes are abrasion and sorting. Collisions of moving particles with each other and with the fixed bed result in abrasion, the process

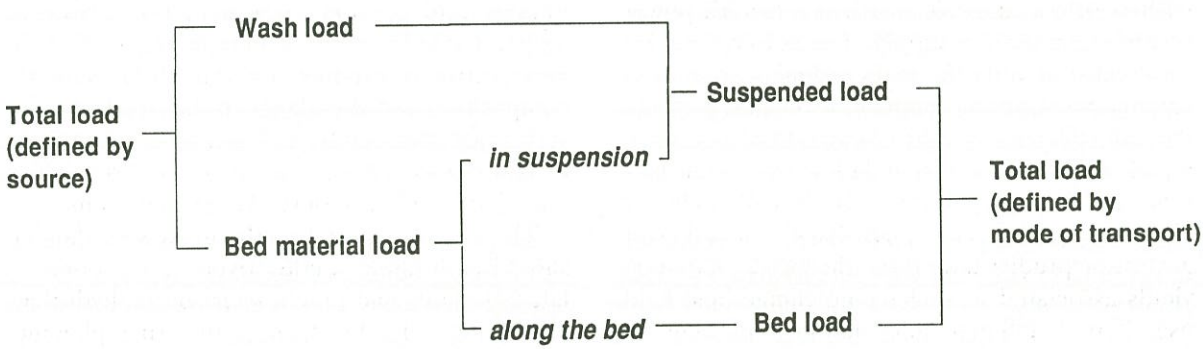

Figure 13.1 Classification of river sediment load by source and mode of transport.

by which the sediment load gradually becomes finer and rounded as grains are split and spalled. For given flow conditions, the intensity of abrasion depends on both the rock type and size of the sediment particles. Clearly, a soft schist will abrade faster than a hard quartzite. Abrasion is much more vigorous during bed-load transport of boulders and gravel; smaller grains of sand size, generally reduced to individual mineral fragments, are more resilient and have fewer collisions as they move in suspension. Whereas abrasion is the process by which a river alters the size of its bed material, differential transport of sediment grains of different mobilities is the process by which a river sorts and spreads its bed material in the downstream direction. The easier, selective entrainment of finer grains thus explains why headwater streams have bouldery poorly-sorted beds and high bed load, while coastal reaches are more likely to have uniformly-graded sandy beds and a dominant suspended load.

While larger velocities generally are required to mobilize larger grains, the cohesion provided by clay minerals can actually result in muds requiring larger velocities to erode than silts or sands. Also, all grains come to rest at velocities less than those required to start them moving. This is because the resisting forces due to friction and cohesion are less for moving than for stationary grains. This effect is particulary striking with fine cohesive sediments which require velocities of about 1 metre/second to be entrained yet deposit in velocities of about 0.1 m/s.

The ultimate "sink" for a river's sediment load is the bed of the ocean - generally near the shore for the sands and gravels, but often spread across the continental shelf or diverted into submarine canyons for the finer fractions. However, sediment is also trapped more or less permanently in lakes and reservoirs, estuaries, floodplains, and the river beds themselves where there is a surplus of sediment supply over transport capacity. An accounting of the supplies of river sediment from its various sources, the transport rates of its bed and suspended loads through the river channel, and deposition rates in temporary and permanent sinks is called a sediment budget. Compiling a sediment budget is an essential part of most river sediment investigations.

Determining River Sediment Load

The requirements for determining the suspended and bed loads differ, reflecting their different origins, transport processes, and distribution within the flow. The concentration of the suspended load, being partly wash load, is related more to the sediment supply than to the river's capacity to transport it and must generally be measured directly. In contrast, the bed load is a function of the transport capacity of the flow and lends itself more to estimation from theoretical or empirical equations.

Suspended Load

The suspended load is measured by first measuring the water flow rate, then collecting samples of the water-sediment mixture. Since the suspended sediment concentration generally varies with depth, position across the channel, and time, a careful sampling strategy and special sampling equipment are required so that the total suspended load in the channel can be accurately determined with a minimum number of samples and effort.

Sediment concentration can vary considerably between the bed and the water surface. The extent of sediment dispersion above the bed depends on the intensity of turbulence, reflected in the upward velocities of turbulent eddies which decay away from the bed, and the velocity at which sediment grains fall through the water - called the "fall velocity" (clearly, grains that fall faster than turbulence can lift them cannot be suspended!). For grains of similar density, fall velocity increases with size. Thus sand grains tend to be concentrated in the zone of most intense turbulence near the bed, while the slower settling silt and clay grains are dispersed more evenly throughout the flow. A typical sediment load, which comprises sand, silt, and clay grades, has a composite concentration profile that decreases away from the bed (Figure 13.2). A specially designed "depth-integrating" sampler is used to average out this concentration profile. A depth-integrating sampler consists of a hollow streamlined brass "bomb" containing a

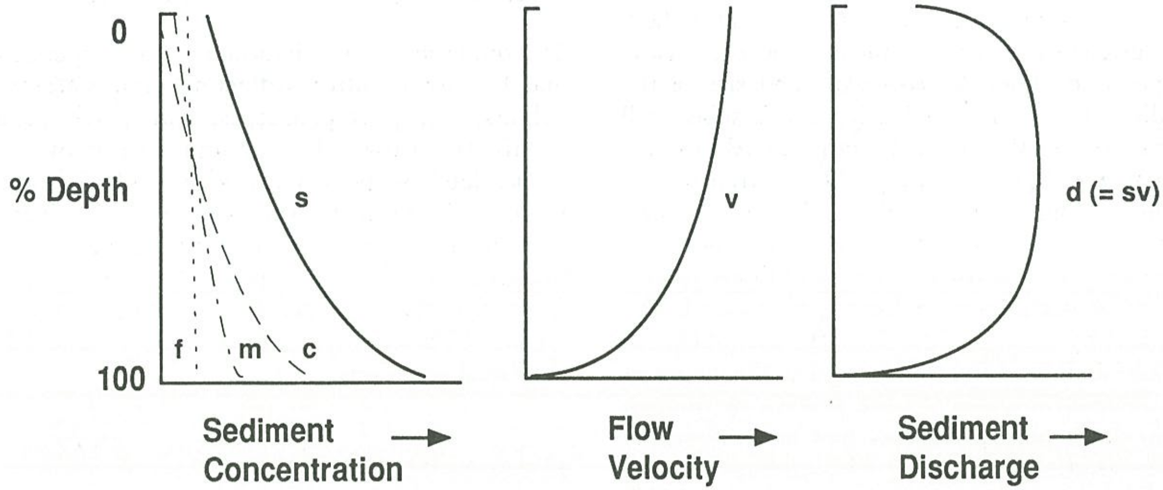

Figure 13.2 Profiles of suspended sediment concentration, flow velocity, and sediment discharge between river bed and surface: "f" is for silt and clay; "m" is for fine sand; "c" is for coarser sand; "s" is composite profile for a mixture of fine-coarse sediment; sediment discharge "d" is the product of the sediment concentration "s" and the water velocity "v".

glass sample bottle, with an intake and air-exhaust system specially designed to ensure that the sediment-water sample enters the sampler at the local stream velocity (Figure 13.3). By traversing the sampler through the flow between the surface and the bed and back again at a uniform rate, a sample is collected whose concentration equals the mean velocity-weighted concentration in the vertical (Figure 13.2).

The actual concentration of sediment in the sample is determined in the laboratory by measuring the volume or mass of the water-sediment mixture, filtering the sample through glass-fibre filter paper, then weighing the dry mass retained on the paper.

Variations in concentration across the flow tend to reflect variations in flow intensity and size distribution of the suspended load. They are allowed

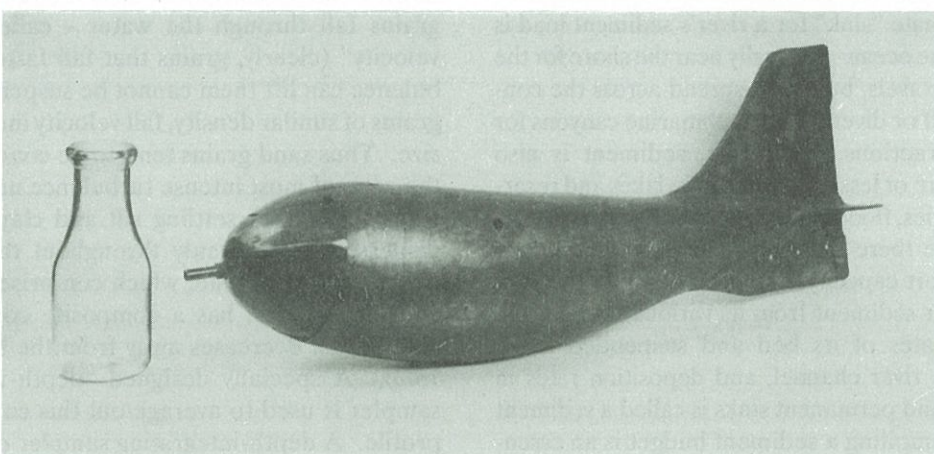

Figure 13.3 US D-49 depth-integrating suspended-sediment sampler and sample bottle.

for by sampling at a number of vertical profiles located at the centroids of sub-sections carrying equal portions of the total water discharge. Fewer sampling verticals are required the more uniform the flow and the greater the proportion of fines in the load. Typically in New Zealand's rivers, samples have been collected from 3 to 5 verticals. The product of the sub-section water discharge and the depth-integrated concentration at the centroid gives the subsection sediment discharge. Sediment discharge summed over all sub-sections, divided by the total water discharge, gives the "mean discharge-weighted concentration" for the whole section - the desired end result of a suspended sediment gauging.

Even so, this result is the concentration for only one instant in time whereas a continuous record is needed to determine the yield of sediment over a period - e.g. a storm or a year - or the average yield over many years. For New Zealand rivers, which often rise and fall rapidly and whose sediment concentrations can vary rapidly, continuous sampling would be time-consuming and expensive. Instead, two other methods of estimating concentrations are used.

The first method involves collecting "index" samples at a single point in the channel cross-section, then converting the point concentration to a mean concentration for the entire cross-section by using an assumed or experimentally determined relationship. The index sample can be collected by hand, but usually an automatic pumping sampler is used. Typically, these can collect 24 or more samples and can be programmed to sample at fixed intervals of time or water level, or at a rate proportional to the flow rate. The main handicap of automatic samplers is their limited number of sample bottles. Alternatively, a turbidity sensor may be used to obtain a continuous trace of suspended load. The point turbidity, which is sensitive to sediment properties as well as concentration, also requires at-a-site calibration with the mean concentration. Turbidity calibration may be difficult where contributions from various sediment sources result in varying sediment properties.

The second method involves compiling "rating" relationships, i.e. determining the usual suspended sediment load for a given water discharge. The sediment rating is then used to convert water discharge records into time-series of sediment discharge or concentration. These sediment ratings can have any arbitrary time base, although normally they are compiled from "instantaneous" values. Owing to the large ranges in sediment concentration and water discharge that occur in rivers, sediment ratings are routinely plotted on log-log graphs (Figure 13.4a). A rating equation of the form

$$C = a\ Q^{b} \qquad (13.1)$$

where C is the sediment concentration, Q is the water discharge, and a and b are coefficients, is fitted either by eye or, more commonly, by least-squares regression of the log-transformed data. With the latter approach, a correction factor may be applied to counteract bias induced by the log-transformation (Ferguson, 1986), although this correction factor should be applied with care (Cohn et al., 1989).

Often, a wide range of sediment concentrations may be observed for a given water discharge, resulting in point-scatter on the sediment rating. Some of this scatter arises from measurement error, but more is due to seasonal effects (e.g. winter snow covering sediment sources in catchment headwaters), whether the water level is rising or falling, antecedent catchment conditions (e.g., a large recent storm may have mobilised much sediment), and to varying sediment sources and times for sediment "slugs" to travel downstream between their sources and the gauging section. In small to medium catchments, commonly much of the available sediment, often stored in or on the banks of the channel, is flushed out early during a storm, while the water level is still rising, resulting in higher concentrations on the rising limb of the storm hydrograph than on the falling limb (e.g. Christian and Thompson, 1978, Figure 13.4b). In larger catchments, this effect may be reversed, complicated, or absent, depending on the proportion of the suspended load derived from the bed-material, and on whether sediment pulses from tributaries arrive at the gauging site before or after the flood peak.

Where some of these effects are identifiable, it is possible to subdivide the set of concentration vs.

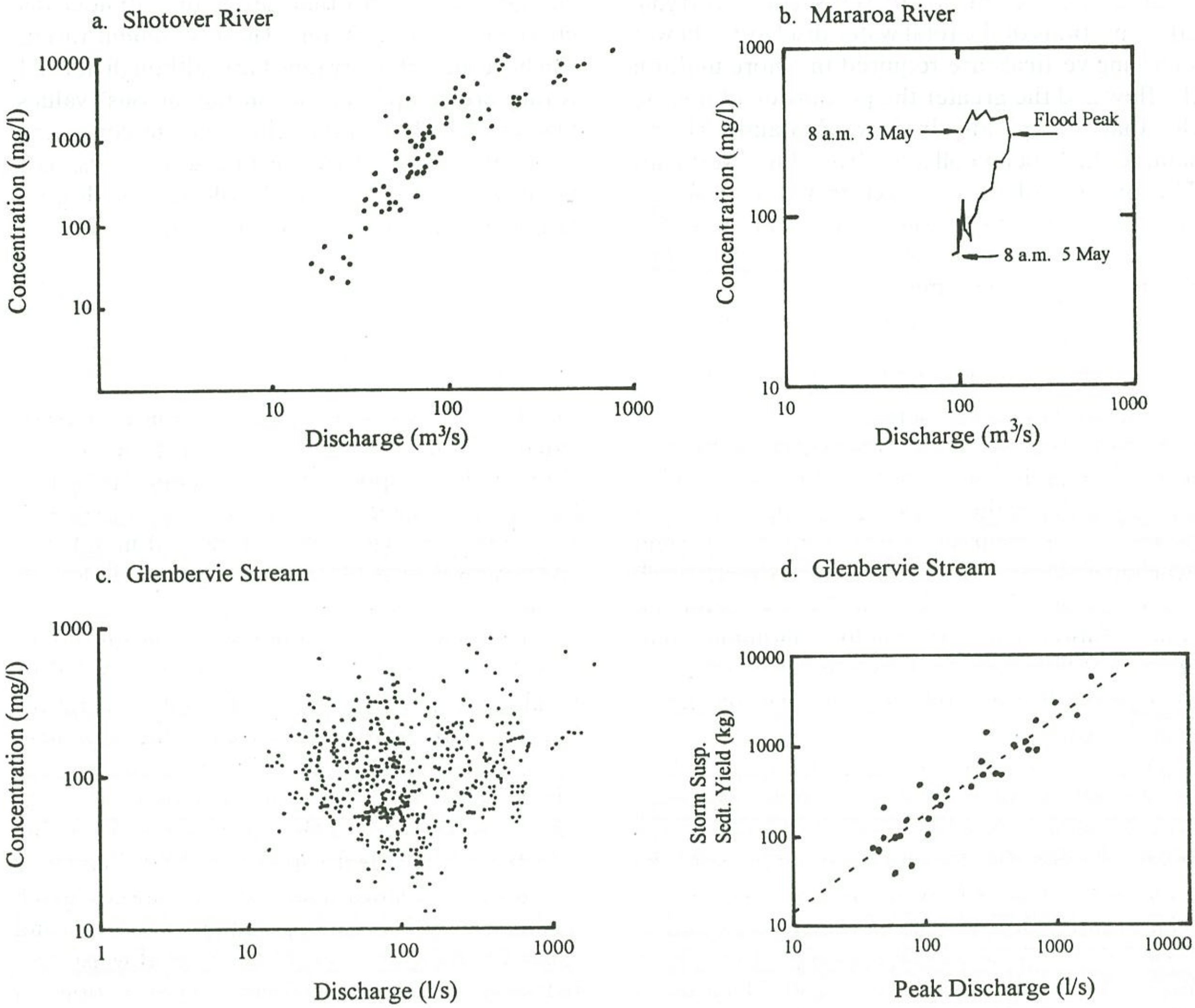

Figure 13.4 Suspended-sediment rating curves: (a) Shotover River - curve typical for a medium to large river; (b) Makaroa River - clockwise-looping of rating during a single flood; (c) Glenbervie Stream - "shotgun blast" concentration rating typical of very small catchments; (d) Glenbervie Stream storm-yield rating.

water-discharge data pairs and produce more accurate ratings (with less scatter) for individual seasons or for rising and falling water levels. When compiling an instantaneous sediment rating curve to be used in estimating the mean annual yield over a number of years, it is important to conduct paired water-sediment gaugings over a wide range both of discharge and of the conditions that cause the rating scatter. In most rivers, the bulk of the sediment yield over a long period is carried by flows ranging between the overall mean flow and the mean of the annual peak flows (i.e., the mean annual flood), and so sampling efforts should be focussed in this range (Figure 13.5).

In small streams, with catchment areas less than a few square kilometres, the instantaneous sedi-

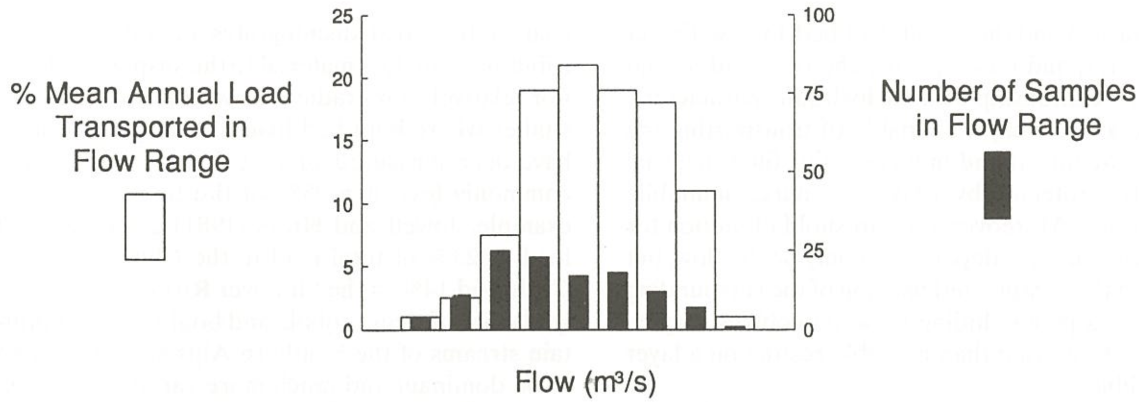

Figure 13.5 Proportion of long-term average suspended sediment load carried by, and number of samples from, given flows - Shotover River. The mean flow of the Shotover River at the sampling site is 37 m^3/s, while its mean annual flood flow is 478 m^3/s.

ment concentration for a given discharge can vary greatly, with concentration responding more to random injections of sediment from erosion sites (e.g., Figure 13.4c). In these cases, the instantaneous rating is rarely adequate. Instead, good relationships can often be found between storm sediment yield, measured accurately for a sample of storms using an auto-sampler, and some index of storm magnitude such as the peak discharge, runoff volume, rainfall, or rain erosivity (Hicks, 1990). Integrating the yield from many sources over a storm appears to average-out many of the random source effects. Once established from a relatively short but intense period of sampling, such ratings (e.g., Figure 13.4d) can be used to estimate the sediment yield over a much longer period of storm runoff or rainfall record.

Bed load

There are four approaches to determining a river's bed load: theoretical formulae; assuming the bed load to be a proportion of the measured suspended load; surveying the accumulation of bed-material in a trap such as a reservoir over a period of time; or direct sampling.

More often than not, because it is easier than direct measurement, bed load is calculated from a formula. Four main classes of formula relate bed-load transport rate to either water discharge (e.g. Shulits, 1934), shear stress (e.g. Meyer-Peter and Muller, 1948), or stream power (e.g. Bagnold, 1980), or else treat bed load transport as a probabalistic process (e.g. Einstein, 1950). The bases of these formulae range from totally theoretical to empirical, although most rely on at least some empirical calibration and so should be applied to only a limited range of conditions. The Meyer-Peter and Muller (1948), Einstein-Brown (Brown, 1950), and Bagnold (1980) formulae have commonly been used for New Zealand's gravel-bed rivers (Carson and Griffiths, 1987).

Bed-load formulae of any type tend to provide better results when applied to individual sub-sections and integrated across-channel than when used simply with hydraulic parameters averaged for the entire cross-section (Carson and Griffiths, 1987). Results are also more reliable when the yield is averaged over a period of time (e.g. Carson and Griffiths, 1989; Young and Davies, 1990). Invariably, formulae predict instantaneous transport rates poorly, tending to overestimate the true bed load. This is because the formulae predict "equilibrium" bed-load transport, which assumes an unlimited supply of bed material, and are generally calibrated from laboratory experiments where equilibrium transport prevailed. "Non-equilibrium" conditions prevail in natural rivers, where the supply (e.g. from the bed, eroding banks, and tributaries) and the mobility of bed-load sediment fluctuates, and consequently the bed load has no unique relationship with the hydraulic parameters. While a river may be capable of transporting the finer fractions found in its bed, this finer material may be protected by a layer of coarse, immobile "armour". Moreover, the threshold of motion for a given grain size depends not only on the flow, but also on the mixture and packing of the surrounding grains - a pebble hiding between cobbles may be harder to entrain than a cobble resting on a layer of pebbles.

In upland rivers, which are characterised by riffle-and-pool morphology, and also in braided channels, the bed load tends to move in waves or "slugs" whose phase and frequency sometimes are, and at other times are not, related to the passage of flood waves (e.g. Hayward, 1979; Griffiths and Hicks, 1980; Beschta, 1983a and b; Ackroyd and Blakely, 1984; Carson and Griffiths, 1989). For example, at times fluctuations in bed load at-a-site appear to relate to the amount of bed material stored in pools upstream, which tend to empty as water level rises and to refill as it falls; on other occasions the stream channel may be swamped with gravel from a few or even a single erosion site.

Thus in natural rivers, particularly ones with gravel beds, where flow is rarely beyond the range of threshold of motion, equilibrium transport formulae should be applied with care, and particular regard should be paid to the range of conditions for which the formulae were derived. In a recent review of transport formulae for gravel-bed rivers, Gomez and Church (1989) find favour with Bagnold's (1980) formula where hydraulic information is limited to section-averaged parameters, but suggest the Einstein (1950), Parker (Parker et al., 1982), and Ackers-White-Day (White and Day, 1982) formulae where information is available on local hydraulics and/or on bed state and grain-size distribution.

Another empirical approach is to assume that the bed-load transport rate, averaged over time, equates to some proportion of the suspended or total load. A figure of 3-10% of total load is often assumed (e.g. Griffiths and Glasby, 1985). However, this proportion tends to vary, generally decreasing downstream due to sorting and as the coarser bed load disintegrates or abrades, contributing more fine material to the suspended load. For relatively low gradient New Zealand channels, studies where both bed load and suspended load have been measured suggest that the bed load is commonly less than 25% of the total load. For example, Jowett and Hicks (1981) estimated bed load as 23% of total load in the Clutha River at Clyde and 14% in the Shotover River). Conversely, studies of steep, cobble and boulder-bed mountain streams of the Southern Alps show bed load to be dominant and much more variable. For example, Hayward (1979) found bed load to be 93% of total load in Torlesse Stream, while Griffiths and Hicks (1980) measured the proportion at 86% in Dry Acheron Stream.

Measuring bed-material accumulations in a trap is generally the most accurate method of determining bed-load transport over a period of years. The usual technique is to monitor deposition in the trap by periodically re-surveying a network of cross-sections. The traps may be purpose built (e.g. Griffiths and Hicks, 1980, Figure 13.6; Cuff, 1981) or deposition may be measured in a fortuitous location such as an hydro-electric reservoir (e.g., Jowett, 1984), an irrigation dam (e.g., Bishop et al., 1984), a river delta or alluvial fan (e.g., Mosley, 1978; Griffiths and McSaveney, 1986), or simply a terminal reach of river channel (e.g. Griffiths, 1979a). Where some suspended load as well as bed load is deposited in the trap, the suspended-load component needs to be subtracted to leave the true bed load input. The size distributions of the suspended load and the trap material are required for this (e.g., Jowett and Hicks, 1981). Some traps may only catch a portion of the load entering them; in such cases the bed-load outflow, and thence the efficiency of the trap, must be estimated by other techniques.

Direct measurement of bed-load transport can be made either with special structures or with bed-load samplers. Structures which divert all of the bed load passing a cross-section into a weighing apparatus, such as the vortex-tube flume on Torlesse Stream described by Hayward and Sutherland (1974) and shown in Figure 13.7, provide a continuous and accurate record of the bed load. They have provided much of the current evidence

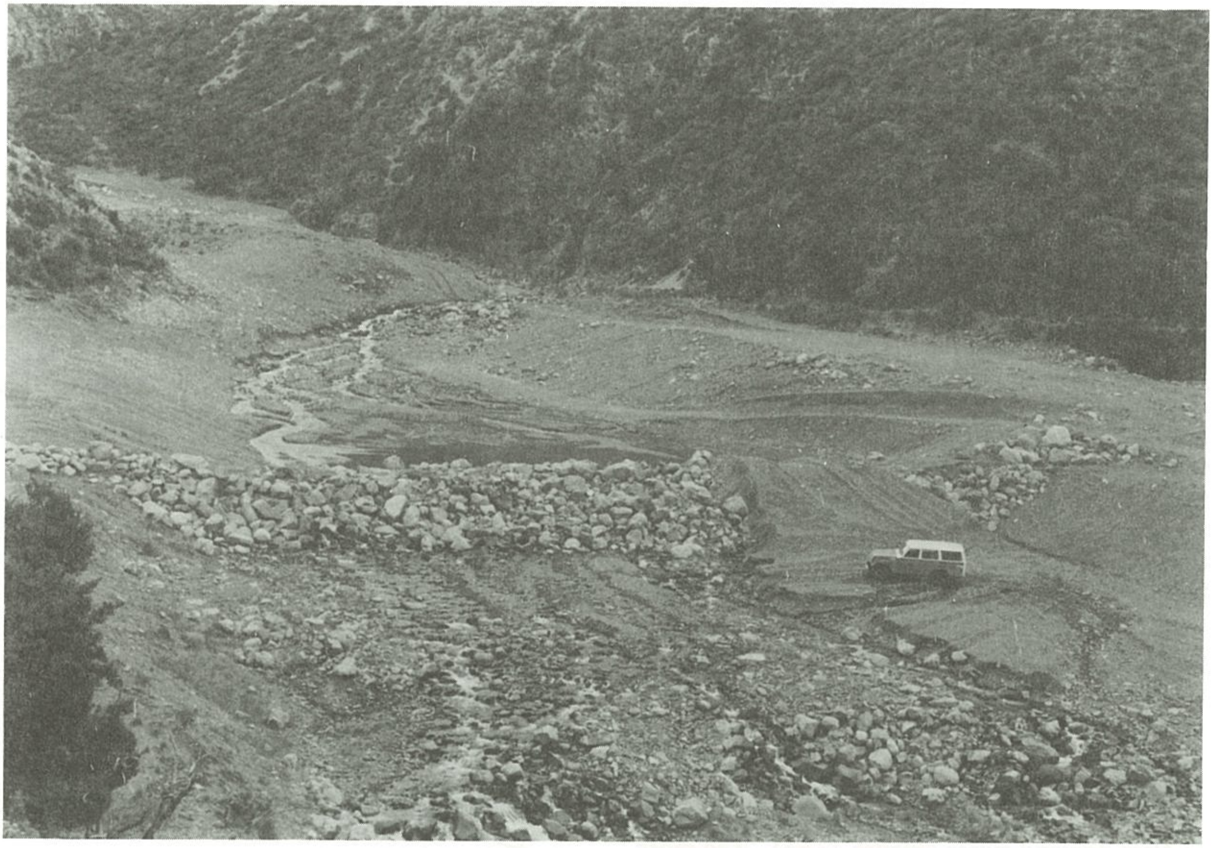

Figure 13.6 Sediment trap formed by permeable rock dam on Dry Acheron Stream, Canterbury foothills (photo I.E. Whitehouse).

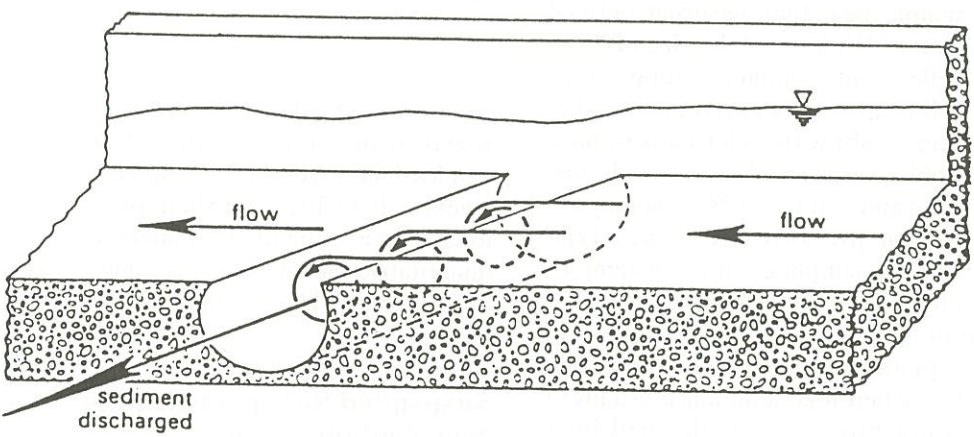

Figure 13.7 Vortex-tube bedload trap at Torlesse Stream, Canterbury (after Hayward and Sutherland, 1974).

concerning fluctuations in the transport rate (e.g., Hayward, 1979), but, unfortunately, they are expensive and limited in scale, and so are suited only to experimental investigations. Collecting discrete samples is the only recourse in larger streams with more intense bed-load transport and for regular monitoring.

Bed-load sampling, however, is far from a straight-forward exercise and requires a sampling strategy tailored to individual sites and flow conditions. Detailed sampling experiments, such as those at Torlesse Stream, have shown that over time-periods varying from a few minutes to several hours, bed-load transport can vary from zero to four times the temporal mean as sediment slugs or bedforms such as dunes or bars pass by. Also, at any given time, the transport rate can vary across the channel by similar amounts, due to spatial variations in the flow velocity and the supply of bed load. Depending on the situation and accuracy desired, up to 40 or more discrete bed-load samples, including repeat samples at each vertical, may be required to average out these fluctuations. Edwards and Glysson (1988) outline methods for bed-load sampling.

Two types of bed-load sampler have been used in New Zealand rivers (Figure 13.8a). The more commonly used sampler is the "Helley-Smith" sampler. This is termed a "pressure-difference" sampler by virtue of its using an expanding intake orifice to overcome the drag developed by the flow through the sample bag, thus ensuring optimal sampling efficiency. With an intake size of 75 or 150 mm, the Helley-Smith sampler is suitable only for sand and fine-gravel bed rivers. Basket samplers, (Figure 13.8b) with wider intakes, have been used in cobble-bed rivers (Davoren and Mosley, 1986; Laronne and Duncan, 1986) but they are less efficient than pressure-difference type samplers. Bed-load sampling with any sampler becomes impractical at high flow velocities, high bed-load transport rates, and large grain sizes.

As with suspended-sediment sampling, the general objective of bed-load sampling is usually to derive a relationship between the bed-load transport rate and the flow rate; this rating can then be used to estimate long-term bed-load yield from flow records (or to verify a relationship deter-

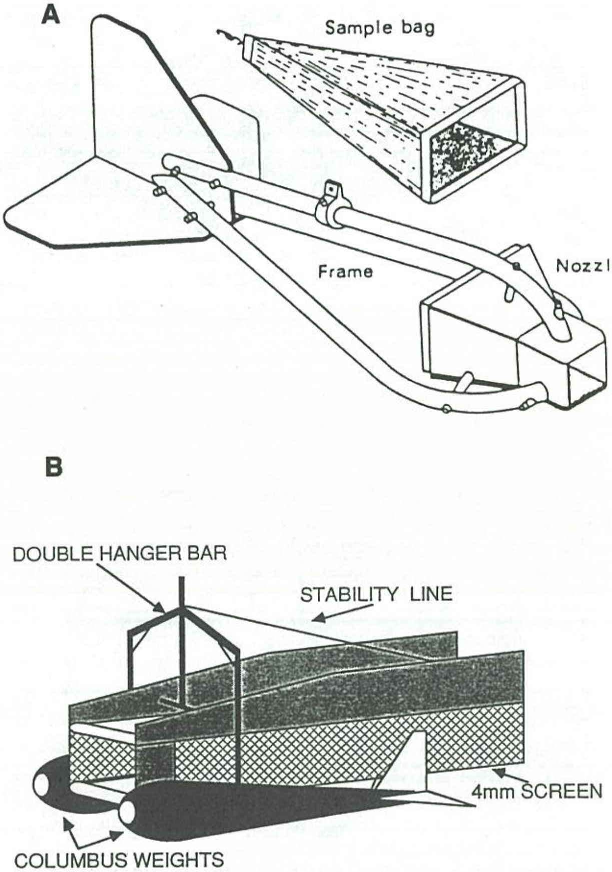

Figure 13.8 Bedload samplers: (a) Helley-Smith pressure-difference type; (b) Davoren basket-type fitted with lead "Columbus" weights.

mined theoretically). Because of the problems and practical limitations discussed above, however, bed load has rarely been sampled in New Zealand rivers. Routine sampling programmes have focussed on suspended sediment, and most of our quantitative knowledge on sediment loads in New Zealand rivers is from this.

Suspended Sediment Load of New Zealand Rivers

Sampling of suspended sediment in New Zealand rivers commenced in 1955 at sites in the central and eastern North Island. This sampling, during high-flood flows when the rivers carry much of their load, was conducted by hydrological survey teams of the Ministry of Works (MOW). Its impetus then was provided by soil erosion and conservation studies. Interest in sediment information for other applications soon led to MOW teams sampling around much of the country, generally at sites in Representative Basins (Ministry of Works, 1970) or in catchments of interest for hydro-electric power development.

The data compiled up to the late 1970's were analysed by Thompson and Adams (1979), Adams (1979 and 1980) and Griffiths (1979b, 1981, 1982) to estimate mean annual suspended sediment yields on regional and national scales. Relatively little information was added to this dataset during the 1980's. Recently, the NIWAR Environmental Data (the descendant of the Ministry of Works hydrological survey) resumed sampling suspended sediment during floods at a selection of sites in their National Hydrologic Reference Network in order to improve the spatial coverage of suspended-load information around New Zealand and clarify the factors causing variations in suspended load from catchment to catchment.

The MOW/NIWAR data are stored on a computer archive maintained by NIWAR Environmental Data in Wellington. It contains information from 389 sites, although only half of these have sufficient data for establishing sediment-rating curves. Further suspended sediment data are found with Regional Councils and research institutions such as the Forest Research Institute.

Together, these data show a considerable range in suspended-sediment concentrations, loads, and particle sizes around the country and at any given site. The concentration at a site depends on the sampled flow conditions. Around the country, peak concentrations during floods range from a few hundred to a few thousand mg/l for relatively small and undisturbed catchments in low hill country (e.g. O'Loughlin et al., 1978 and 1984; Dons, 1988). In the larger rivers draining the eastern Southern Alps, concentrations are higher but rarely exceed 12,000 mg/l. Surprisingly, concentrations measured in the energetic, steep South Island west coast rivers rarely exceed 6,000 mg/l, perhaps because the sediment entering their channels is quickly flushed by the frequent floods and never builds up to allow larger concentrations. The highest concentrations have been measured in the Papa mudstone catchments of the East Cape region of the North Island. There, concentrations of 40-60,000 mg/l are not uncommon, and a concentration over 90,000 mg/l has been measured in the Waiapu River. High concentrations, of the order of 30,000 mg/l, have also been measured in small catchments disturbed by earth-moving machinery, such as during forest harvesting or urbanisation.

Much less information has been collected on particle-size distribution of suspended sediment; available data show that it can vary widely, depending on flow conditions, the nature of the bed-material, and the geology of the catchment. For example, streams and rivers draining catchments with glaciers or with bedrock that is either soft mud-siltstone or deeply weathered tend to have a suspended load dominated by clay-silt (i.e, wash load), while rivers with sand and sandy-gravel beds, such as those draining the schist catchments of Otago and the greywacke catchments of Canterbury, tend to have significant sand fractions also (Figure 13.9).

Adams (1979, 1980), Griffiths (1979b, 1981, 1982), and Thompson and Adams (1979) showed how the sediment yield, generally compared in terms of the mean annual yield per unit catchment area (termed "specific yield") varies around the country. These authors all used the sediment rating technique and essentially the same set of MOW concentration and flow data. Consequently, in most instances their estimates agreed within the limits permitted by the scatter on the sediment-rating curves, and in a few cases they were verified with independently derived information on yields from surveys of sedimentation rates in reservoirs.

The specific yield data of these authors show enormous variability (Table 13.1). Intermontane regions of Otago and Canterbury and low rainfall regions of the North Island had specific yields ranging from 30 to 100 t/km^2/yr, whereas yields commonly were several hundred t/km^2/yr over much of the North Island and the coastal foothills of the eastern South Island. Yields range from one

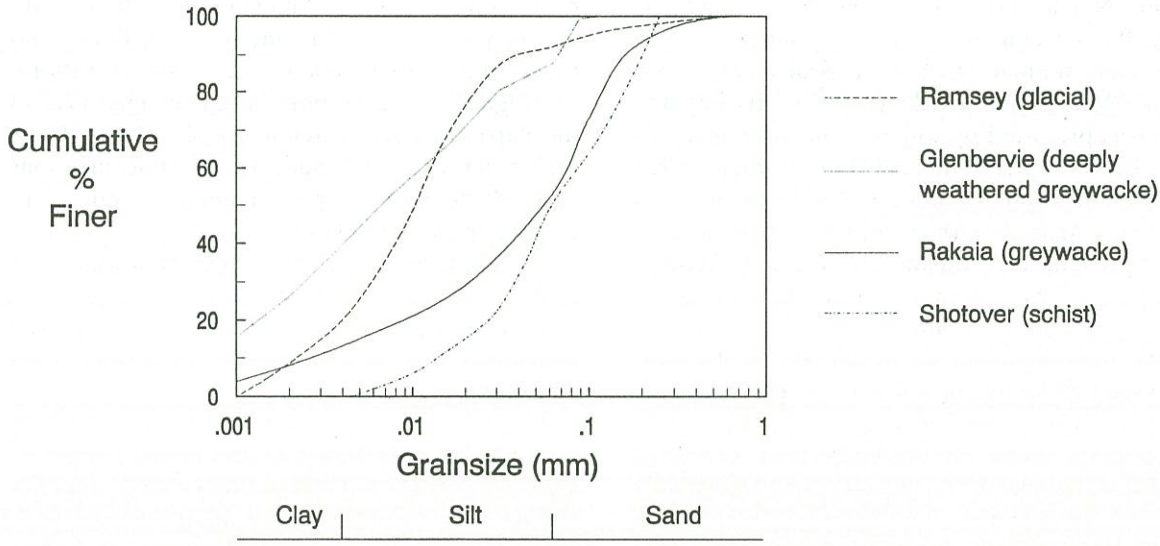

Figure 13.9 Size distributions of suspended load in some New Zealand rivers.

to several thousand t/km^2/yr for catchments draining the eastern Southern Alps from the main divide and the Ruahines, to 10 to 30,000 t/km^2/yr for schist catchments of Westland spanning the zone of intense rainfalls (up to 10-12 m/yr annual precipitation) west of the main divide and for the mudstone catchments of the North Island East Cape region. The highest measured specific yield in the North Island is 19,970 t/km^2/yr from the Waiapu Catchment (Griffiths, 1982), while the highest measured specific yield in the South Island is 29,600 t/km^2/yr from the Cropp River, a schist catchment located in the highest rainfall area of the Southern Alps (Griffiths and McSaveney, 1983). These yields rank among the highest in the world (Griffiths, 1979b).

Griffiths (1981, 1982) found that catchment precipitation was the main factor affecting specific yield. On this basis, he subdivided the country into eight separate sediment-yielding regions (four in the North Island, four in the South) and presented for each regression equations relating suspended sediment specific yield to mean annual precipitation (Figure 13.10). Griffiths and Glasby (1985) used existing sediment yield data plus estimates of the sediment yield from unmeasured catchments, based on these regional equations, to compute the total input of river sediment to the New Zealand coast (they assumed bed load was always 3% of suspended load). They estimated the yield to the South Island coast was 284 ± 40 million tonnes/year and to the North Island coast was 105 ± 9 million tonnes/year. 75% of the South Island yield is delivered to the west coast, while 69% of the North Island yield is transported to the east coast between Gisborne and East Cape.

River sediment yields can change in response to changes in climate, land use, or river diversion and damming. The existing MOW/NIWAR data have not been explored for such trends, nor would any be likely to show clearly owing to the relatively short and often patchy records at most sites. Qualitative changes in sediment yield, however, have been observed over historic times associated with land-use change. Sediment loads in some rivers were much higher than their present values during the phases of accelerated hillslope erosion that followed widespread forest clearance and tussock-land burning (Whitehouse, 1984). The evidence for this includes direct observation and records of sedimentation in traps such as floodplains and estuaries (e.g., Hume and McGlone, 1986; Hume and Gibb, 1987). Moreover, recent studies in small experimental basins indi-

Region	River	Catchment Area (km^2)	Main Rock-type	Precipitation (mm/yr), Suspended Sediment Yield (t/km^2/yr)	Reference	
Bay of Plenty	Motu	1393	Greywacke & argillite	2690	1961	Griffiths (1982)
	Whakatane	1557	Sandst & siltstone	1600	236	Griffiths (1982)
N. Is. East Coast	Waiapu	1378	Sandst & mudstone	2400	19970	Griffiths (1982)
	Hikuwai	307	Sandst & mudstone	1900	13890	Griffiths (1982)
	Waipaoa	1582	Sandst & siltstone	1600	5836	Griffiths (1982)
	Tukituki	2380	Sandst & siltstone	1500	445	Griffiths (1982)
	Esk	254	Sandst & siltstone	1500	1096	Griffiths (1982)
Central-West N. Is.	Waitara	725	Sandst & siltstone	2200	644	Griffiths (1982)
	Wanganui	6643	Sandst & siltstone	1800	326	Griffiths (1982)
	Mangakino	373	Ignimbrite	1400	35	Griffiths (1982)
	Tongariro	772	Greywacke & andesite	2000	555	Griffiths (1982)
Southern N. Is	Raumahunga	2340	Sandst & siltstone	1400	247	Griffiths (1982)
Eastern S. Is.	Rakaia	2640	Greywacke & argillite	3000	1641	Griffiths (1981)
	Waimakariri	3210	Greywacke & argillite	1900	1669	Griffiths (1981)
	Waiau	1980	Greywacke & argillite	2000	1300	Griffiths (1981)
	Selwyn	164	Greywacke & argillite	1300	584	Griffiths (1981)
S Is. Inland Basins	Ahuriri	557	Greywacke & argillite	1600	98	Griffiths (1981)
	Manuherikia	2036	Schist	830	35	Griffiths (1981)
	Pomohaka	1924	Schist	930	29	Griffiths (1981)
	Shotover	1088	Schist	1600	1019	Griffiths (1981)
S. Is. West Coast	Buller	6350	Granite	2600	270	Griffiths (1981)
	Grey	642	Granite	3000	552	Griffiths (1981)
	Hokitika	352	Schist	9400	17070	Griffiths (1981)
	Haast	1020	Schist	6500	12736	Griffiths (1981)
	Cropp	29	Schist	10070	29600	Griffiths & McSaveney (1983)
	Ivory	2.1	Schist + glacier	8600	14900*	Hicks et al. (1990)

* Total sediment yield

Table 13.1 Mean annual suspended sediment yields and other characteristics of selected New Zealand river basins.

cate that sediment yields may surge up to 100-fold in the years following forest clearing (O'Loughlin et al., 1980; Hicks and Harmsworth, 1989). The instigation of soil conservation legislation and measures in the 1930's-1940's largely contained the historically elevated sediment yields, at least on broad, regional scales, and since then most New Zealand rivers probably have maintained a reasonably stable sediment load regime.

Grant (1985) interpreted dated sequences of cut and fill terraces in river valley bottoms as evidence of changes in sediment regime over much of New Zealand during the last 1800 years. He ascribed cycles of erosion and sedimentation to a climate irregularly fluctuating between warm stormy conditions, which led to increased erosion and sediment transport, and cooler calmer conditions, which led to soil re-formation and less sediment transport. A matter of concern to us all is how climate change associated with current global warming might alter river sediment regimes over the next century or two. A change towards warmer conditions with more frequent and intense rainstorms, and consequently greater sediment transport, is considered possible for the north-easterly regions of New Zealand (Salinger and Hicks, 1990; Basher, 1990; Griffiths, 1990).

Factors Controlling River Sediment Yield

Sediment yield is influenced by many factors, including precipitation, runoff, geology, soil type, vegetation, land use, glaciation, and topography. Research to date suggests that the main factor

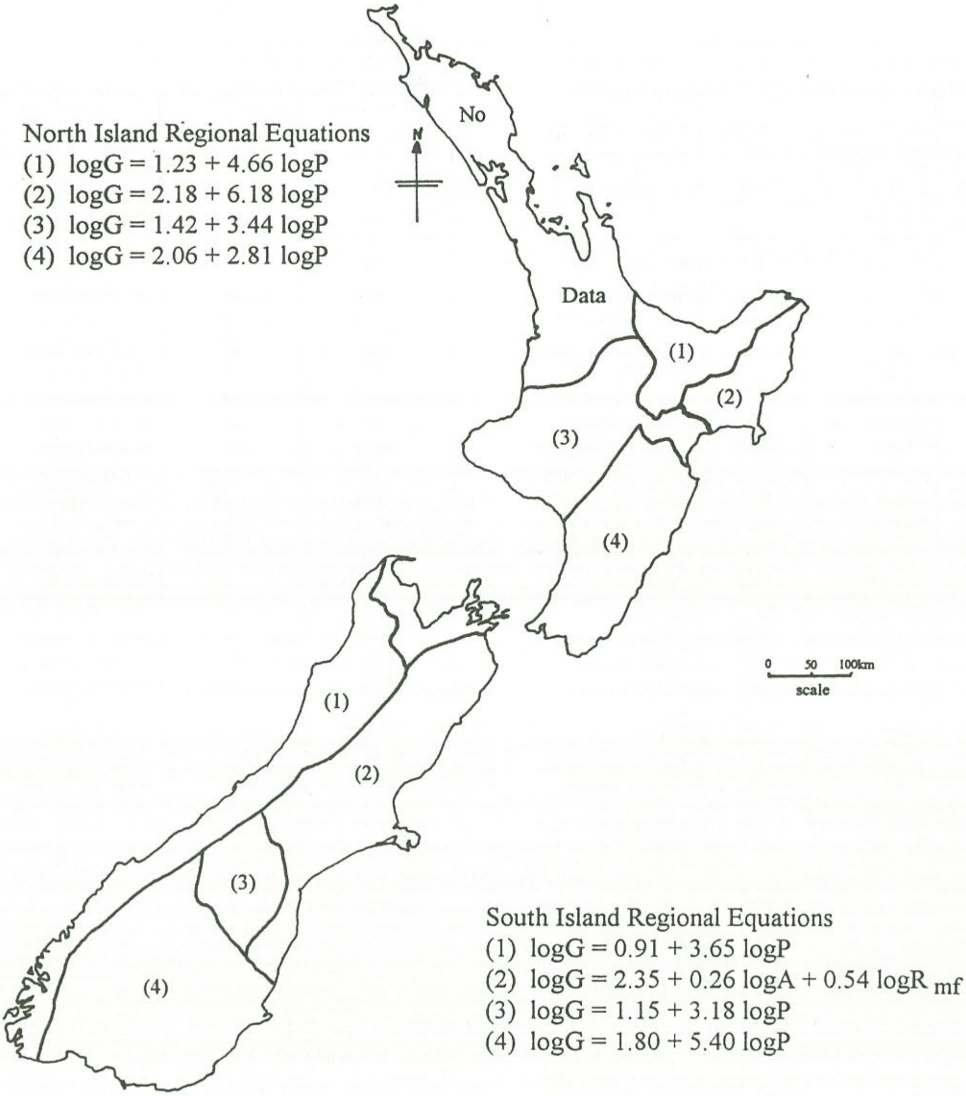

Figure 13.10 New Zealand sediment regions and regional sediment yield equations of Griffiths (1981, 1982). G is suspended sediment specific yield ($t/km^2/yr$), P is precipitation (m/yr), A is catchment area (km^2), and R_{mf} is runoff from mean annual flood ($m^3/s/km^2$).

causing variation in specific yields among New Zealand rivers is precipitation - both the total amount and how it falls through the year. Catchment geology and land use can explain much of the residual variation. The extent of glacial cover, in some Southern Alps catchments, has no noticeable influence.

Precipitation and Runoff

Griffiths (1981 and 1982) demonstrated the dominant influence of mean annual rainfall (or its surrogate, mean annual runoff, which is fed by rainfall) on variation in specific sediment yield among New Zealand rivers. This should be no surprise given that a greater rainfall should both erode more sediment and produce more runoff to carry that sediment through the drainage network, and since there is a wide variation in rainfall over the country, ranging from less than 1 m/yr in South Island intermontane basins to over 12 m/yr in the western Southern Alps.

Griffiths also suggested that specific yield among basins with similar annual rainfalls should vary with the intensity, frequency, and duration patterns of storms. He showed that yields were higher where rain fell in high-intensity, short-duration events, rather than in prolonged low-intensity events. This is because, compared to prolonged drizzle, cloud-bursts of the same water yield have greater erosive potential and produce more intense and greater runoff. This has been confirmed by Hicks (1990), who found at small catchments in various parts of the country that sediment yields from individual storms correlated better with a rainstorm erosivity parameter than with the storm rainfall. By summing the square of the incremental rainfall over the storm, Hicks' erosivity parameter gives added weight to more intense storms.

Extreme rainstorms, with return periods ranging from decades to centuries, can also influence the mean annual sediment yield. Such events often appear to surpass a threshold condition, "shocking" the catchment, generating intense hillslope erosion, and injecting vast quantities of sediment into the river channel. For some years after these events, the sediment yield may be abnormally high; then, over years or decades, as the hillslopes restabilise and the sediment slug is flushed through or stabilised in the river channel, the yield may decay exponentially to a "normal" regime until the next major event occurs, completing a geomorphic cycle (see O'Loughlin et al., 1982; Beschta, 1983a and b; and Grant, 1985 for New Zealand examples; also Schumm, 1979; and Kelsey, 1980). Cyclone Bola, which struck the North Island East Cape region in 1987, may be the most recent example of such a "threshold" event in New Zealand (Figure 13.11).

Figure 13.11 Erosion scars caused by Cyclone Bola, Lake Tutira, Hawkes Bay (photo N. Trustrum).

Geology and Glaciers

The influence of catchment geology on sediment yield is made readily apparent by the extremely high specific yields from the East Cape region (Table 13.1). There, soft Tertiary mudstone and active tectonism result in steep, easily eroded hill-slopes, large supplies of fine-grained sediment to the river channels, and specific yields of the order of 25 times those of North Island catchments with similar rainfall regimes but harder rock types such as greywacke (Griffiths, 1982).

A geological influence is also suggested in the variation of specific yield in the Southern Alps (Figure 13.12). There, at least where mean annual precipitation is less than about 4 m, schist basins yield about 10 times more sediment per unit area than do basins with harder rocks such as granite, gneiss, and limestone. This can be explained by the fissile nature of the schist and its composition of silt and fine-sand grade minerals, which permit its relatively easy disintegration into the size fractions readily transported as suspended load.

The influence of glaciation on sediment yield in the Southern Alps was investigated by Hicks et al. (1990). They found no detectable difference in specific yield between glaciated and unglaciated schist basins receiving similar precipitation, and suggested that in the Southern Alps environment, the sediment yields were determined more by the intensity of the processes of erosion and sediment transport than by the type of process. Their study of glaciated Ivory Basin showed that much of its sediment yield originates from non-glacial erosion

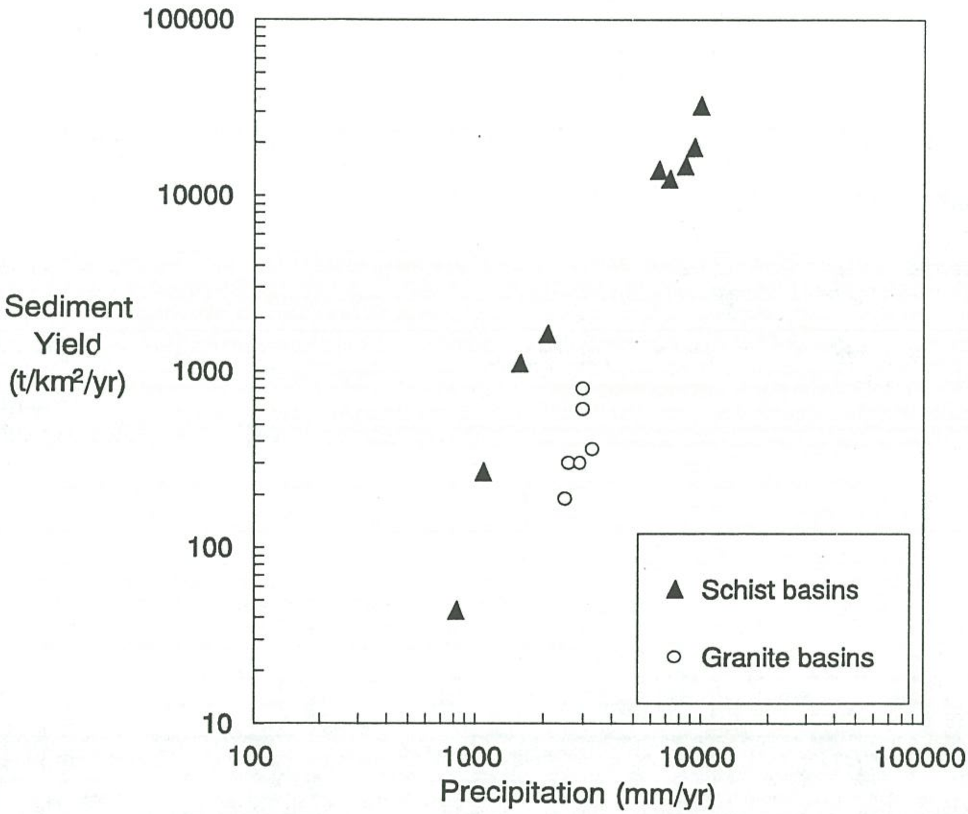

Figure 13.12 Mean annual sediment yield vs. mean annual rainfall for schist and granite catchments in the Southern Alps and Westland.

of its steep rock walls, and that this slope sediment is readily sluiced through and beneath the glacier by rain and melt-generated runoff.

Land use and Land-use Change

Land use, generally represented by vegetation cover, exerts further control on sediment yield. While the land use is stable, its influence is generally much smaller than that of precipitation. Consequently, the effect of land use has had to be demonstrated by detailed paired-basin studies in which two nearby basins, differing only in terms of their land use, are monitored concurrently. Hicks (1990) compared storm sediment yields from 12 small basins (areas less than 3 km^2) that were either in pasture or exotic forest. He found that pasture basins yield more suspended sediment than do exotic forest basins, except in regions lacking frequent high-intensity rainstorms. Overall, his pasture basins yielded 1.6 to 4.5 times more sediment than the exotic forest basins. Dons (1987), in a study comparing basins near Rotorua, found too that exotic forest basins yield less sediment than matched basins in pasture, and that a native forest basin yielded least sediment of all. O'Loughlin et al. (1984), from another paired-basin study in inland Otago, found no noticeable difference in sediment yields from tussock and burnt-tussock catchments. The lack of large storms over the two-year study period, however, makes this result inconclusive.

In contrast to the relatively modest influence of established land use on sediment yield, changes in land use can induce major, but usually temporary, changes in sediment yield. Conversion of forest or tussock-land to grazing land is a prime example. Evidence, mainly qualitative, abounds as to the effects of historical forest clearing and tussock burning, subsequent over-grazing and rabbit plagues, and associated increased hillslope erosion on high inputs of sediment to river channels, both in the South Island high country (e.g., Whitehouse, 1984) and the North Island hill country (e.g., Schouten, 1977).

Studies by O'Loughlin and Pearce (1976) and O'Loughlin et al. (1980) in north Westland show that sediment yield increases up to 100-fold following clear-felling of native forest. Most of the extra sediment originates from logging roads and from slope failure due to loss of tree-root reinforcement of soil strength. Five years after logging, the high sediment yields begin to wane; they subsequently return to "normal" when exotic forest is established. Hicks and Harmsworth (1989) calculated that the several-year period of harvesting activity at Glenbervie Forest, an exotic plantation in Northland, provides approximately 70% of the sediment yield over the 32-year growing-logging cycle.

Urbanisation induces similar large surges in sediment yield (e.g., Herald, 1989). While earthworks may quickly be paved or built over, years may pass before sediment slugs are flushed from the stream channels.

Direction of Future Sediment Load Investigations

While we currently have an outline sketch of the sediment load transported in New Zealand rivers, much remains to be known if future sediment management issues, be they concerned with channel engineering works, environmental impacts, or the damage wrought by extreme events like Cyclone Bola, are to be handled in the most practical and cost-effective manner. There are two main areas where future investigations of sediment load ought to be focussed.

The first involves further sampling programmes directed at: (i) extending the spatial coverage of basin sediment-yield information; (ii) clarifying the roles played by rainstorm characteristics, geology, land use, and land-use change on sediment production; (iii) improving estimates of the yield and size-distribution of sediment delivered to reservoirs and coastal beach systems; and (v) quantifying the temporal variability and extremes of suspended-sediment concentration.

The second involves process studies and modelling, particularly of transport of gravel bed load in mountain streams and braided rivers. The challenge is to improve first the conceptual, then the hydraulic models of the bed-load transport process to the stage where reliable forecasts can be made of medium-term average and extreme loads. Only then will we be on a sound footing to predict the changes to river channels induced by catchment or channel works, extreme natural events such as cyclones and earthquakes, and changes in climate.

Acknowledgements

We thank I.G. Jowett and M.J. McSaveney for comments on this manuscript.

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14
River Hydraulics and Instream Habitat Modelling for River Biota

Ian Jowett

Introduction

The flowing water of rivers transports sediment, nutrients, aquatic and terrestrial food for stream inhabitants, and provides passage between habitats occupied by different species and their life stages. The living organisms in a river, together with their environment, form an ecosystem. The concept of an ecosystem is a broad one, and emphasises the relationships between species and between the species and the environment, as well as the energy transfers involved in supporting the system. These interrelationships are important when predicting biological changes likely to arise from any change in the physical stream environment. The physical stream environment is in turn influenced by the nature of the catchment and river banks as well as by the quantity and variability of river flow. Water velocity is the primary characteristic that distinguishes rivers from other aquatic freshwater ecosystems such as ponds and lakes. Water velocity influences the physical structure of a river, the habitat it provides, and the biota that live in it (Jowett and Duncan, 1990).

The Food Chain

The transfer of energy from the primary producers, plants, through a series of organisms that feed on plants or other organisms, is known as the food chain. In a stream community, plants (principally periphyton or attached algae in New Zealand rivers (Biggs, 1990)) rely on sunlight and nutrients in the water for growth. The benthic invertebrates (insects living on the stream bed) that graze on periphyton, such as some mayflies (e.g. *Deleatidium* spp.), beetles (*Hydora* spp.) and chironomids, are amongst the most common in New Zealand rivers (Quinn and Hickey, 1990a). Benthic invertebrates form the major portion of the diet of many of the 27 native and 20 introduced freshwater fish species (McDowall, 1990). Thus environmental changes that affect the periphyton community can in turn alter the composition and abundance of benthic invertebrates and hence fish.

River Structure and Hydrology

The shape and form of a river is largely the result of erosion and deposition sediment. Both the morphology of the river and sediment transport depend on the nature of the riverbed and banks, the gradient of the channel, and the flow. These factors govern the velocity of water in the channel, which in turn determines the ability of water to move or erode the stones or soil – the process of shaping a river. However, in most New Zealand rivers, constantly varying flows and a plentiful supply of sediment from a rapidly eroding landscape tend to disturb any equilibrium between water and the riverbed and to influence the biota that can colonise the river.

Rivers can be divided into zones of differing character and biotic composition along their length. In the headwater zone, river gradient is steep, average water velocity fast, substrate large, and water temperatures are low. As a river flows from its headwaters to the sea there is usually a longitudinal reduction in gradient, water velocity, and substrate size and an increase in water temperature. All these factors influence the abundance and composition of the biota of a river. The physical characteristics of a river, and the amount and duration of the river flow, can affect fish passage. Passage from one part of the river to another or to the sea is essential for the life cycle of many New Zealand fish species. Trout, especially rainbow trout, migrate to upstream spawning areas in the winter months (Hicks and Watson, 1985). Adult eels migrate from the rivers to sea to spawn. Inanga (a common galaxid species) spawn in or near estuaries; when their eggs hatch, the young are washed into the sea where they feed and grow, and subsequently return to the rivers as whitebait. Other galaxids, some of the bullies, and torrent fish have similar sea-going life cycles and only a few of the galaxid and bully species reside in the river for their entire life.

Flow Variability

In New Zealand rivers, flow varies with the intensity and frequency of precipitation and to a lesser extent with the ability of the surface soils and lakes to store water, and with catchment size and slope (Jowett and Duncan, 1990). Flow variability is thought to influence the number of plant and animal species present in a river. Resh et al. (1988) postulate that a stable system would be dominated by the superior competitor or the one most suited to the environment; an unstable system by the most efficient colonisers, and an intermediate system by a combination of those species. Jowett and Duncan (1990) showed how the biota in New Zealand rivers vary with flow variability. East coast rivers which experience extended periods of drought and infrequent but large floods, tended to have high concentrations of ions and nutrients (Close and Davies-Colley, 1990) and biota that prefer low water velocities. In contrast, rivers with little flow variability, such as those fed by frequent rainfall or from groundwater storage, had higher average water velocities, low concentrations of ions and nutrients, and were dominated by "clean water" flora and fauna (Biggs et al., 1990).

Although many animal species can usually respond to flow changes by moving to more suitable locations, the total population of a river system probably responds on a longer time scale. Jowett (in press) showed that the abundance of adult brown trout was more highly correlated with the amount of habitat available at mean annual low flow than the amount available at either median or mean flow – flows which occur more frequently than the mean annual low flow.

Flow Extremes

Extreme flows often affect the abundance of stream biota. Trout numbers, especially of small trout, are usually severely reduced by floods (Allen, 1951; Jowett and Richardson, 1989). Algae and aquatic macrophytes can be scoured by floods and aquatic invertebrate abundance and community composition can change, although invertebrate populations tend to recover quickly (Sagar, 1986; Scrimgeor et al., 1988; Biggs and Close, 1989; Quinn and Hickey, 1990b). During droughts, low levels of dissolved oxygen and high water temperatures can kill trout and community composition changes to that which favours low water velocities. Some biologists believe that such disturbances, particularly floods, are the dominant organising factor in stream ecology (Resh et al., 1988). Some species will not be able to adapt to harsh environmental conditions, such as drought or flood, whereas those better adapted to the environment and with short life cycles will be able to recolonise quickly after such events.

Water Temperature and Quality

Water temperature limits the distribution and rates of growth of many species. Trout and some species of benthic invertebrates such as stoneflies and mayflies (Jowett, 1990; Quinn and Hickey, 1990a) are found in colder waters, whereas other species, such as mosquito fish, are found only in warmer waters (McDowall, 1990). One introduced fish species, the sailfin molly, is found only in one thermal swamp in the central North Island (McDowall, 1990). The temperature of flowing water is most affected by exposure to incoming radiation and ambient air temperature, so there is a general correlation between mean annual water temperature and site elevation and latitude (Mosley, 1982a). The amount of shade can significantly influence water temperature (Williamson et al., 1990) and the effects can be modelled (e.g. Theurer, 1982).

Most New Zealand rivers have high quality water with a pH of 7.0-8.5 (slightly alkaline), low turbidity in baseflow conditions (Close and Davies-Colley, 1990), and oxygen levels at or near saturation. All rivers sampled in the "100 rivers" study had dissolved oxygen levels in excess of 6 g/m^3 (Close and Davies-Colley, 1990). Although many stream animals are susceptible to reduced oxygen, problems tend to occur only in polluted streams where bacterial decay reduces oxygen levels, or in rivers with warmer water, low flows, and macrophyte beds, such as those on the east coasts or in the north of New Zealand. In the latter group of rivers, the oxygen requirements of the macrophytes during the night, combined with high temperatures and low aeration rates, can reduce oxygen to low levels at dawn (Close and Davies-Colley, 1990).

Physical Character of Flowing Water – River Hydraulics

Water Velocity, Depth, and Substrate

Many stream insects use the current to assist respiration and for the supply of food (Hynes, 1970). Some species, such as stoneflies and net-spinning caddisfly, have very specific velocity, depth, and substrate requirements, whereas the distribution of others, such as cased caddisfly, appears to have little relationship to water depth or velocity. Water velocity is the "driving force" of a stream ecosystem, broadly determining the distribution and community composition of periphyton, benthic invertebrates, and fish.

Water depth also influences the distribution of stream biota. Adult trout are often found in the deeper areas, whereas juvenile trout, invertebrates and periphyton are usually more abundant in shallow water. Adult trout select deep water for shelter or cover and low water velocities, whereas juvenile trout are small enough to use the cover provided by gravel and cobbles in faster-flowing runs or riffles. Periphyton requires light for growth and thus is more abundant in shallower waters with high nutrient levels.

Most benthic invertebrates live on or among the stones of the stream bed and feed on periphyton growing on them or on detritus trapped there. Shallow, fast-flowing water with coarse substrate provides prime habitat for these types of invertebrate. Substrate can also provide shelter for fish and mobile invertebrates. A substrate of sand or soft silt generally supports only a small number of species, usually in low densities.

In natural rivers water velocity, depth, and substrate are interrelated. In most rivers the size of the bed material increases with water velocity because the force of the current winnows out finer particles. Velocity, and often substrate size, decreases as depth increases in rivers with well-defined pool-riffle sequences. However, in more uniform rivers the reverse occurs; with the main current in the deeper section of the river and shallow slow-flowing water at the margins. In biological studies these interrelationships make it difficult to separate the different effects of velocity, depth, and substrate on biota.

Vertical Velocity Distribution

Stream invertebrates and most New Zealand fish species are usually more common on or near the river bed and are influenced more by water velocities in this region than the mean water velocity. The maximum water velocity normally occurs between the surface and 0.25 of the depth below the water surface and is about 1.2 times the mean velocity (Figure 14.1). The mean velocity is usually measured at 0.6 times the depth below the surface, but in deeper water a better measure of the mean is obtained by averaging velocity measurements taken at 0.2 and 0.8 of the depth. Equations for the velocity at any point in the vertical can be derived theoretically. The velocity at any point (v) at a height y above the stream bed is given by:

$$v/V_m = 1 + (7.83n/Y^{1/6})(ln(y/Y) + 1)$$

where V_m is the water mean velocity, n is the Manning roughness parameter, and Y the total depth. This equation does not apply very near the bed and at the surface. An alternative approach is to take measurements of both mean water velocity and velocity at the depth of interest, often 0.1m above the river bed and develop an empirical regression equation of the form $v = V_m(a(y/Y)^b)$, where a and b are constants (Stalnaker et al., 1989).

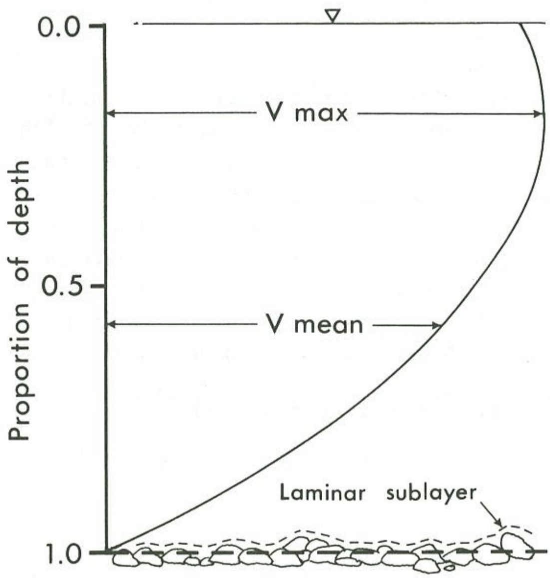

Figure 14.1 Vertical distribution of water velocity in turbulent flow across a rough stream bed.

Flow types and turbulence

Flow in an open channel is classified as either laminar or turbulent depending on the value of Reynolds number. Reynolds number ($Re = V_mY/kinematic\ viscosity$) [1] relates inertia forces to viscous forces and so is important when considering the fluid forces on small objects such as benthic invertebrates. As Re increases, inertia forces increase until instability develops and flow in open channels becomes turbulent at Re of over 500 – a value exceeded at most locations in rivers (Chow, 1959). The total force on an object in flowing water is comprised of a viscous (surface drag) and pressure component – the ratio depending on Reynolds number and body shape. Blunt objects are subject to mainly pressure forces whereas the force on well-streamlined bodies is mainly viscous.

Biologists tend to characterise portions of a river as riffle, run, or pool. Riffles are shallow swift flowing areas with a broken water surface and larger substrate (gravel, cobbles, or boulders). Pools are deeper slow flowing areas with an almost level, smooth water surface and often containing deposits of finer substrate (sand or gravel). Runs are intermediate between pools and riffles and are characterised by an undulating but relatively unbroken water surface. This classification is related to the Froude number ($Fr = V_m/\sqrt{(gY)}$, where g is the acceleration due to gravity 9.81 m/s^2). The Froude number relates inertia forces to gravity forces and is important hydraulically wherever gravity dominates (e.g. waves and open channel flow) and is the criterion that distinguishes tranquil and rapid flow. Tranquil or sub-critical flow occurs where $Fr < 1$ and rapid or super-critical flow where $Fr > 1$ (the "stoppers" of a canoeist). Measurement of typical velocities and depths in pools, runs, and riffles by Allen (1951) and Mosley (1982b) suggests a hydraulic definition of pool, run, and riffle based on Froude number (Figure 14.2). This hydraulic definition of habitat correctly identified the habitat type of 56% of 1112 measurements of velocity, depth, and

(1) The kinematic viscosity of water is 10^{-6} m^2/s at 20°C and decreases about 2% for every °C rise in temperature

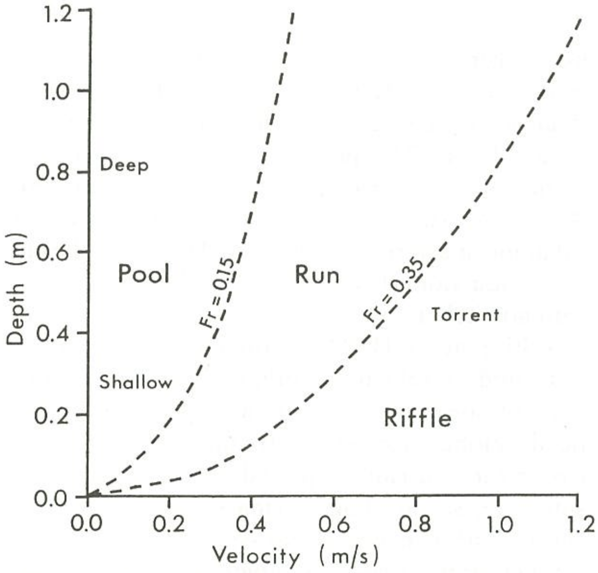

Figure 14.2 Froude number classification of stream habitat types.

habitat type in the Ashburton River (Jowett in prep.). When the water surface slope was taken into account as well as Froude number, the habitat type of 66% of the measurements was correctly identified. Froude number differentiated pools from runs and riffles whereas slope differentiated runs from riffles. Pool habitat was characterised by a Froude number of less than 0.18 and runs and riffles by a Froude number greater than 0.18. Locations where $Fr > 0.18$ and slope > 0.0099 were classified as riffles and locations where Fr > 0.18 and slope < 0.0099 were classified as runs. Consideration of velocity, depth, substrate size, and relative roughness did not improve the ability to identify correctly the habitat type.

Shear stress and boundary layer

Shear stress or drag is the force exerted on the stream bed by flowing water and will also be a measure of the force on benthic invertebrates, periphyton, and aquatic macrophytes. Shear stress on a stream bed is difficult to measure directly. However, the average value of the shear stress on the bed must resist the downslope component of the gravitational force on the water.

Shear stress $\tau = \gamma RS$, where γ is the specific weight of water (density $\rho \times g$), R the hydraulic radius (cross-section area/wetted perimeter), and S the slope. The shear velocity is a hydraulic index derived from the shear stress. It has the dimensions of velocity but is not related to any physical characteristic of flowing water other than shear stress.

$$\text{Shear velocity } U^* = \sqrt{(\tau/\rho)} = \sqrt{(gRS)} = V_m/(2.5\ln(12Y/k_s))$$

where k_s is the bed roughness of the substrate. Bed roughness was originally defined by Nikuradse as the diameter of packed sand grains. Thompson and Campbell (1979) suggest that the roughness of a graded substrate might be about 4.5 times the median substrate size. This is similar to using the D90 or D85 size (the sieve size which passes 90% or 85% of the substrate).

At the stream bed, the water velocity is zero but it rapidly increases with distance from the bed. This layer where velocity changes quickly with height above stream bed is called the boundary layer. In most natural channels there is a thin laminar sublayer at the stream bed (Figure 14.1) above which the flow is turbulent. The thickness of the laminar sublayer Δ is given by:

$$\begin{aligned} \text{Delta} &= 11.6 \text{ x kinematic viscosity/shear velosity} \\ \Delta &= 11.6 \times 10^{-6} \times 2.5\ln(12Y/k_s)/V_m \end{aligned}$$

If the irregularities in the stream bed are small they are totally enclosed by the laminar sublayer and the stream bed is said to be hydraulically smooth. However, if the irregularities are large enough a stable laminar sublayer will not form and the stream bed is said to be hydraulically rough. The transition from smooth to rough occurs at the critical roughness (k_c) (Chow, 1959).

$$\begin{aligned} k_c &\approx 100 \text{ x kinematic viscosity/shear velocity} \\ &\approx 100 \times 10^{-6} \times 2.5\ln(12Y/k_s)/V_m \\ &\approx 8 \text{ x thickness of the laminar sublayer} \end{aligned}$$

The average thickness of the laminar sublayer at over 300 locations in the Clutha, Mangles, Mohaka, and Waingawa Rivers was about 0.2 mm and was disrupted by the substrate (ie hydraulically rough) at 99% of these locations.

Habitat Suitability and the Instream Flow Incremental Methodology

The concept of habitat suitability is all around us. All life has adapted to a particular range of habitats. The concept of "good" habitat is familiar to most people. It is commonly used by anglers and hunters seeking their prey. In the stream environment, habitat usually refers to the physical conditions – water velocity, depth, nature of stream bed, and perhaps the amount of shelter afforded by the banks or substrate. Habitat suitability curves describe what is considered to be "good" habitat. If the range of suitable habitat for a species or life stage of a species can be determined, it is possible to quantify the area of suitable habitat available within a river. This area is termed the usable area. Waters (1976) introduced the concept of weighting the usable area according to degree of habitat suitability, where the habitat suitability could vary from one (optimum) to zero (unsuitable). Once habitat suitability curves or criteria are defined they can be applied to habitat survey data (Figure 14.3) and the amount of suitable habitat calculated. This is the basis of the instream flow incremental methodology (IFIM) (Bovee, 1982; Jowett, 1982). However, appropriate habitat suitability curves should be used. There are significant differences between the brown trout preference curves derived in New Zealand (Jowett, in press) and in the U.S (Raleigh et al., 1984a). The primary requirements for salmonids are both space and food (Chapman, 1966). Assessing instream flow needs for a river based on salmonid space requirements and not considering the production of food is like designing a house with no kitchen.

While most IFIM applications have concentrated on velocity, depth, and substrate, there is no reason why habitat suitability curves cannot be developed for any instream feature or use. Cover for fish can be provided by water depth, substrate, stream banks, riparian vegetation, or surface turbulence. The assessment of such widely varying forms of cover is subjective and there is a danger of duplication if water depth, velocity, and substrate suitability are evaluated independently. The evaluation of the suitability of water depth could include its use as cover and similarly with substrate. Surface turbulence is a result of the combined effect of water velocity, depth, and substrate and could be combined into a hydraulic index such as Froude number if relevant. Although most published IFIM studies have considered fish or benthic invertebrates, the technique can be ap-

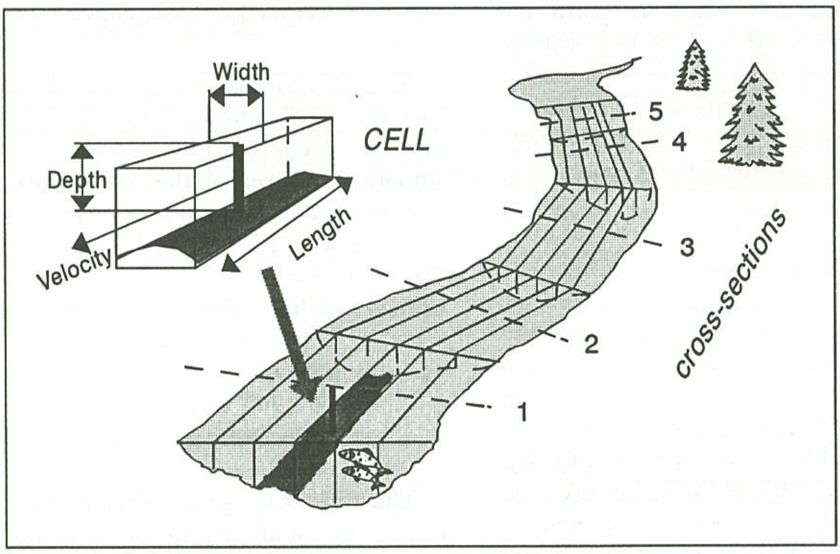

Figure 14.3 Instream habitat model of river reach showing cross-sections and cells.

plied to any stream use. For example, Mosley (1983) lists river width, water velocity and depth requirements for 14 recreational activities ranging from paddling to power boating and water skiing.

Periphyton

The growth of periphyton on the stream bed is closely related to water velocity. Biggs (pers. comm.) has shown that periphyton growth is at a maximum where water velocities are in the range of 0.35 – 0.6 m/s, particularly in low nutrient streams, where low velocity areas such as runs and pools often contain little periphyton growth compared to the riffle areas. High velocities promote growth by increasing the rate of nutrient supply. However, as communities grow during low flow periods, the cells begin to extend upward into regions of higher velocity. When the drag on the cells exceeds their tensile strength, strands of algae are broken off or torn from the substrate. This limits the build-up of periphyton.

Periphyton communities in high velocity water are largely dominated by taxa which are strongly attached to substrate, whereas low velocity areas contain communities with loosely aggregated cells and weak attachments to substrate (Biggs, 1990).

Benthic invertebrates

Stream benthic invertebrates are influenced by water velocity, depth, and substrate. Some stream insects are found only in swift flowing areas, others in slow flowing areas, and some are ubiquitous (Figure 14.4) (Jowett et al., 1991). Insects are usually most abundant in areas with a cobble substrate but some species, such as net-spinning caddisfly, seem to prefer areas with larger more stable substrate, and some, like some species of chironomid, are also abundant in areas with fine substrate. Depth preferences are more difficult determine, but abundance usually decreases with depth. The interrelationship between velocity, depth, and substrate makes it difficult to identify the causal factor. Although insect abundance is very variable, there are significant correlations with habitat suitability based on water velocity, depth, and substrate (Table 14.1).

More complex hydraulic indices, such as Froude number, Reynolds number, shear **stress,** and boundary layer thickness, may also be correlated to insect abundance (Statzner et al., 1988; Jowett et al., 1991). Most mayfly, stonefly, and caddisfly species are more abundant in the shallow fast-flowing water of riffle habitat than in runs or pools. For this reason, the abundance of these

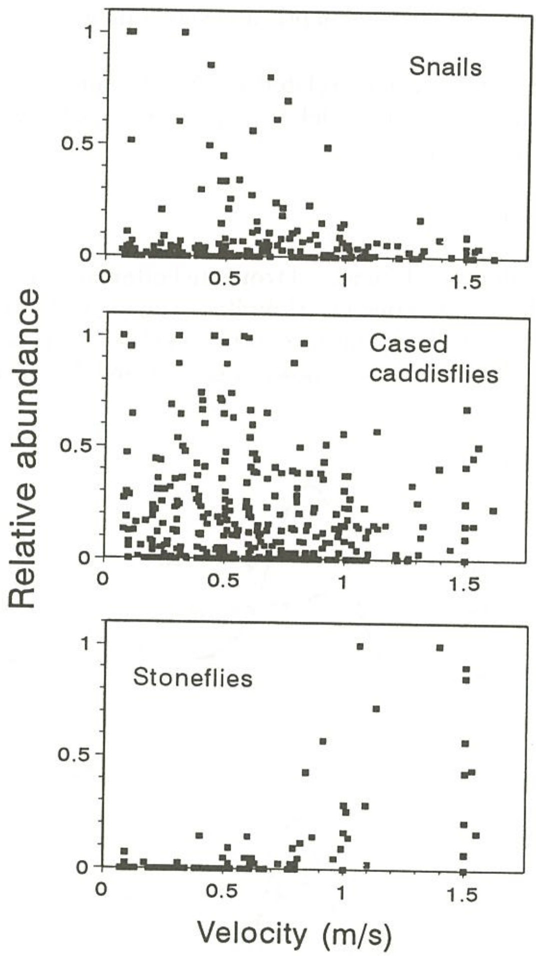

Figure 14.4 Variation of relative abundance of three benthic invertebrate species with water velocity, where each point represents the number of insects in the sample relative to the highest number sampled in the river [Jowett et al., 1991].

Species	River			
	Clutha N=105	Mangles N=68	Mohaka N=80	Waingawa N=81
Mayfly *(Deleatidium spp.)*	0.50	0.45	0.68	0.32
Mayfly *(Coloburiscus humeralis)*	0.53	0.80	0.75	–
Stonefly *(Zelandoperla spp.)*	–	0.68	0.50	0.58
Cased caddisfly *(Olinga feredayi)*	0.17	0.15	0.51	0.11
Swimming mayfly *(Nesameletus spp.)*	–	0.05	0.43	0.50
Fly *(Aphrophila neozelandica)*	–	0.78	0.57	0.54

Table 14.1. Rank correlation coefficients between invertebrate abundance and habitat suitability based on water velocity, depth, and substrate at the sampling point.

insects is usually related to Froude number, the hydraulic index which best differentiates between these habitat types.

Trout

Habitat preferences of trout are better known than those of benthic invertebrates. Angling texts from the turn of the century describe likely trout streams and more recent books (e.g. Hill and Marshall, 1985) accurately describe locations where trout are likely to be found. In New Zealand the physical characteristics of feeding locations used by large brown trout were measured in two rivers. Similar water velocities were utilised by trout in both rivers although the availability of these were different in each river (Figure 14.5). These data can be formed into habitat suitability curves (Figure 14.7) which are, by chance, similar to curves for rainbow trout in the United States! (Raleigh et al., 1984b). In 59

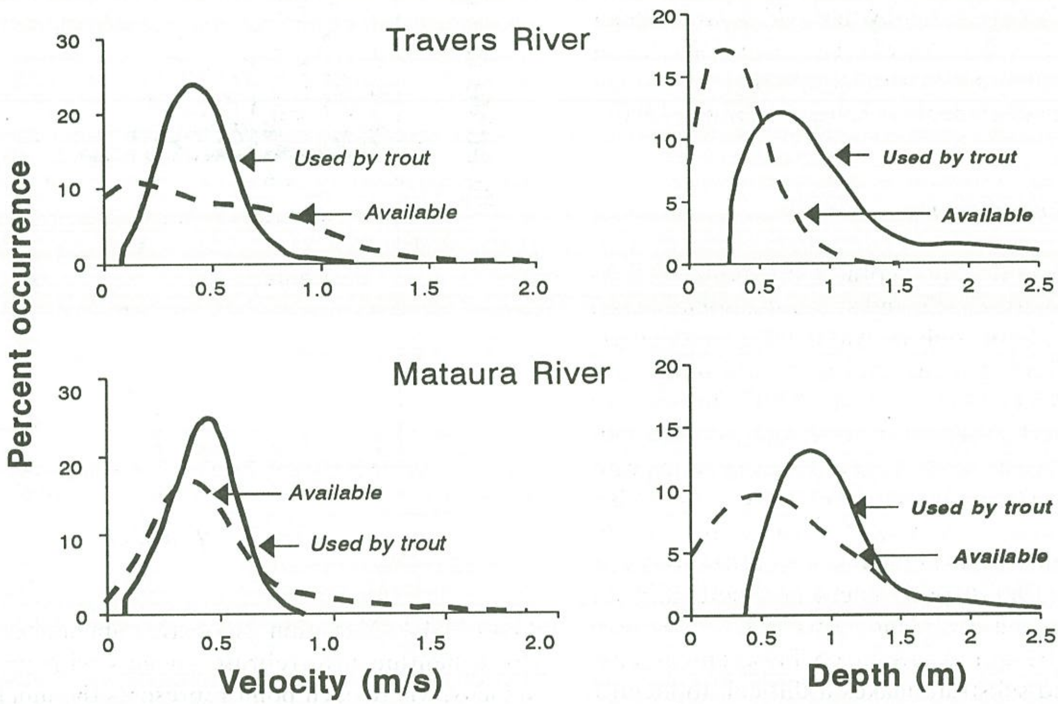

Figure 14.5 Comparison of water velocities and depths available in the Mohaka and Travers Rivers with those used by drift-feeding large brown trout.

New Zealand rivers (Jowett, in press), it was found that adult brown trout were more abundant where there was a high percentage of suitable feeding habitat.

Species may have different habitat requirements at different life stages . Trout spawning requirements are different from the requirements of good adult habitat, and habitat suitable for rearing of fry is different again. Most streams contain a mix of conditions and can cater for a variety of requirements. For instance, trout fry often utilise the slow flowing margins of a stream, and adults the deeper, swifter mid-stream waters. Adult trout are often found in pools, and suitable spawning conditions can often be found at the tails of pools.

Hydraulic Modelling and Prediction of Habitat Suitability

Hydraulic modelling

Flow in an open channel is classified as (1) uniform or non-uniform, and (2) steady or unsteady for the purposes of hydraulic modelling (Chow, 1959). Uniform flow, where there is no change in velocity with distance, rarely occurs in natural rivers, although for practical purposes, flow is considered uniform if depth is approximately constant along the channel. Flow is considered steady if the hydraulic characteristics of flow remain constant over the time interval under consideration. Unsteady flow occurs where the velocity at a point changes with time, such as in surges, bores, or flood waves.

The standard step method, used to model non-uniform steady flow in natural rivers, is well established in engineering practice (Henderson, 1966) and should be sufficiently accurate for ecological applications in most natural rivers. For a given flow, the longitudinal water surface profile can be calculated from the slope, friction, and hydraulic radius (i.e. Manning equation) using principles of energy conservation. The data required are:

- ☐ discharge
- ☐ water surface elevation at the downstream end
- ☐ cross-section area and hydraulic radius at all cross-sections over the range of expected elevations
- ☐ average hydraulic roughness and other causes of headloss (expansions, obstructions, and contractions) between cross-sections.

If water surface elevation at the downstream cross-section is unknown it can be estimated by beginning the computations some distance downstream or alternatively from critical and uniform flow formulae based on the form of hydraulic control at the downstream section. Hydraulic roughness and losses are best determined from field data on discharge, cross-section area, hydraulic radius, and slope. Hydraulic roughness (Mannings n) can vary with flow in an unpredictable manner (e.g. Hicks and Mason, 1991) and this limits the range of flows for which the roughness calibration is valid. Cross-section spacing should decrease where water slopes and cross-section areas change rapidly and increase where hydraulic conditions are uniform. In practice, this means that cross-sections are more closely spaced where habitat types change, such as at the heads and tails of riffles. This procedure is consistent with those used to sample instream physical habitat. Calculations begin at the downstream cross-section and progress upstream. If the starting elevation is incorrect, successive upstream calculations tend to converge on the correct elevation.

Alternatives to hydraulic modelling are available. The simplest method is to measure water levels at all cross-sections over a range of flows. Hydrological techniques, such as fitting an equation to at least three level-flow measurements, can be used to extrapolate to higher or lower flows.

The distribution of water velocities across a cross-section can be calculated once the water level and flow are known (Figure 14.6). Cell velocities can be adjusted for site specific features such as an upstream obstruction which might cause a reduction in velocity or a current on a bend increasing local velocities.

Calculation of habitat suitability

For every cell in the river the suitability of the velocity, depth, and substrate is evaluated on a

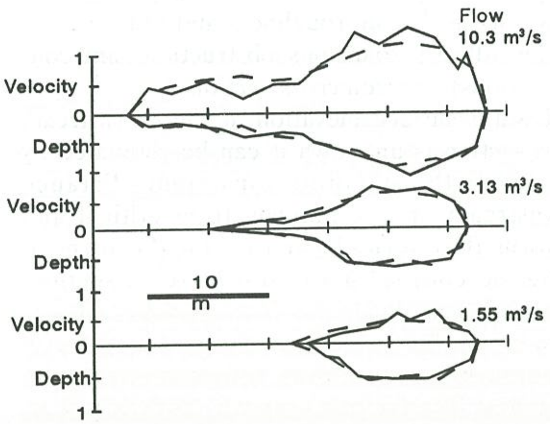

Figure 14.6 Comparison of measured water velocities and depths across a channel (solid line) with those predicted from measurements taken at a flow of 13.2 m^3/s (dashed line) [Mosley and Jowett 1985].

scale of 0 (unsuitable) to 1 (optimum) (Figure 14.7). The point suitability is multiplied by the plan area of the cell it represents (Figure 14.3) and summed over the reach to give the weighted usable area (WUA). Once a hydraulic model of the reach is derived, water velocities and depths can be predicted for any flow and the amount of suitable habitat at that flow evaluated (e.g. Figure 14.8). This provides useful information on the availability of instream habitat and its variation with flow. Computer programmes that combine hydraulic simulation and evaluation of habitat suitability are available (e.g. Bovee, 1982; Jowett, 1989).

Only if the amount of available habitat is limiting population abundance will there be a relationship between habitat and abundance; other factors, possibly unrelated to streamflow, might be limiting the population.

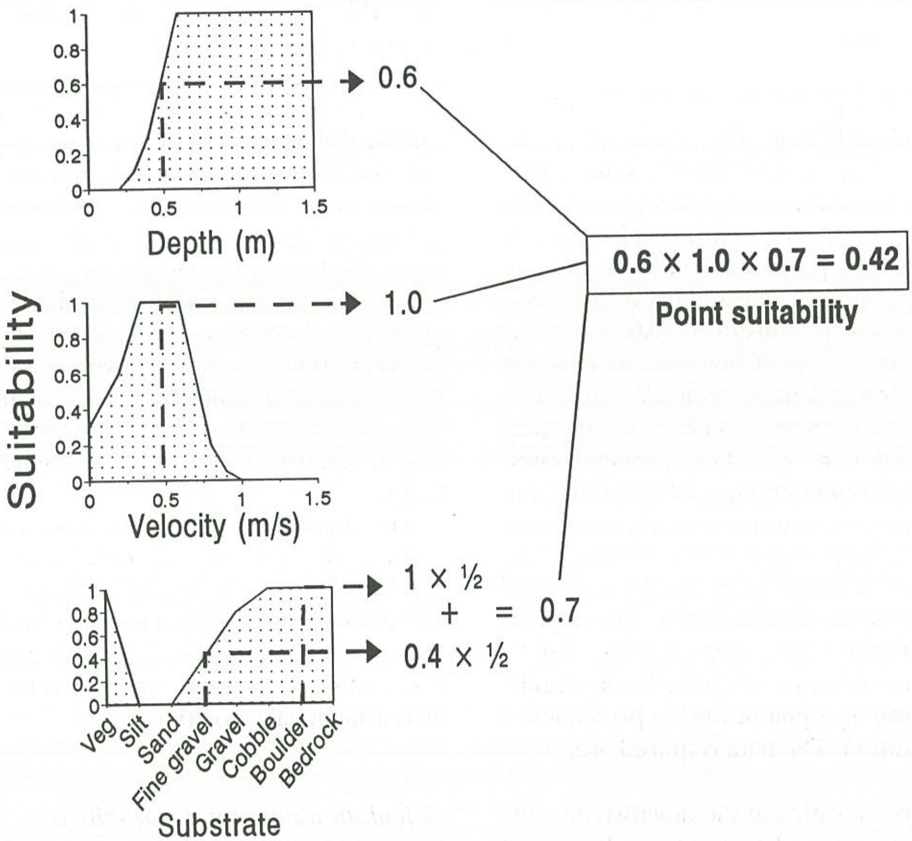

Figure 14.7 Use of habitat suitability curves to calculate point suitability for a water depth of 0.5 m, velocity of 0.5 m/s, and substrate composed of 50% boulder and 50% fine gravel.

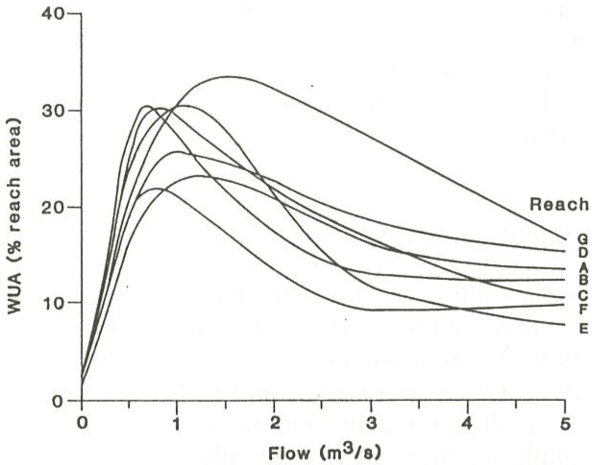

Figure 14.8 Variation of weighted usable area with flow for 7 reaches in the Esk River [Jowett, 1986].

Biological Modelling

Knowledge of the factors controlling populations and the development of biological models which can predict population response to changes in the environment are the next stage in improving water management tools. The development of most models is hampered by the lack of knowledge of the factors controlling the abundance of a species. Even for salmonids, one of the most studied fishes, little is known of what actually controls fish numbers. Many models seem to bear little relationship to the known biology and life cycle of the species. Frequently, it is claimed that the complexity of the interactions and the myriad of factors that affect the species make it too difficult to predict the results of an environmental change.

But is it? Brown trout are found in most New Zealand rivers from the middle of the North Island to the bottom of the South Island. Water temperature is the most likely reason they do not occur further north. Egg mortality begins to increase when water temperatures exceed 11°C (Scott and Poynter, 1991) and brown trout are absent from rivers where temperatures in winter (when trout spawn) exceed about 10.5°C (Jowett, in press). However, water temperature is only one factor which limits the distribution and abundance of trout. Where temperatures are suitable, the presence and abundance of trout will depend on other factors, such as the availability of suitable food, space, and spawning grounds; but, if water temperatures are too high a trout population will not be self-sustaining because of high incubation mortality.

Because stream invertebrates form a major component of trout food, trout food and its variation with flow can be estimated from invertebrate habitat suitability indices. Adult trout habitat can be estimated in a similar way so that the variation of both "food" and "space" with flow can be predicted. These two variables (WUA for "food" and "space") plus seven other variables explained 87.7% of the variation in numbers of large brown trout in 59 rivers (Figure 14.9). The most important variables were **adult trout habitat, food production, instream cover,** and **water temperature** as an overriding factor. The other variables described catchment and stream conditions which can also influence trout abundance. **Sand substrate** is very poor food producing habitat and brown trout are rare in areas where the predominant substrate is sand. **Lake outlets** are well known for their high trout stocks and abundant invertebrate fauna,

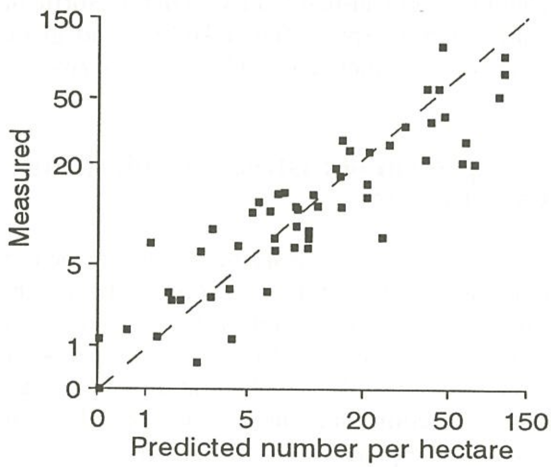

Figure 14.9 Comparison of measured and predicted numbers of large brown trout per hectare in 59 New Zealand rivers [Jowett, in press].

headwaters usually contain lower trout densities than the lower reaches of a river, and high **gradient** rivers are severely depleted by floods (Jowett and Richardson, 1989). **Pastoral development** also appears to have an adverse impact on trout and invertebrate abundance (Quinn and Hickey, 1990b).

In a natural river, flow and the quality of instream habitat vary with time. Trout numbers were much more closely related to the amount of adult trout habitat at mean annual low flow than the amount available at higher flows. This suggests that the quality of habitat at low flow is one of the limiting factors in the system. Similarly trout numbers were more closely related to the amount of habitat for food production at median flow than the amount at either low or mean flow. Thus, even if there is adequate habitat at low flows, a trout population is controlled by the average food producing capacity of the river rather than the capacity during more extreme events.

The rate of growth of periphyton is controlled by the amount of scour during floods, by water velocity and by factors such as nutrient levels, temperature, light availability, and invertebrate grazing (Biggs, 1987). Using three of these variables – days since the last flood, conductivity, and water temperature – Biggs (1988) developed a regression model that explained 76% of the variation in periphyton biomass for 88 samples from a selection of gravel-bedded rivers in the North and South Islands. Spring-fed, lake-fed, and glacial rivers were not included in his sample of rivers.

Management of Instream Habitat and River Ecology

Models such as the brown trout model can be used to predict flow requirements for brown trout. One-dimensional water temperature models can predict the effect of flow changes on water temperature. Adult trout habitat at low flow and food production under normal flow must be maintained at adequate levels. Other factors which influence trout abundance alter little with flow. Cover can vary with flow, but unless the flow modifications are so severe that stream banks are dewatered, there is usually little change.

The periphyton model could also be used to predict the onset of "nuisance" algal growths. For instance, at conductivities of 10, 15, 20, and 25 mS/m, and a water temperature of 18°C, "nuisance" growths could be expected 149, 104, 72, and 47 days, respectively, after a flood (Biggs, 1988).

Habitat quality guidelines

In the "100 rivers study" the habitat quality in 65 rivers around New Zealand was evaluated (Jowett, 1990). In this group of rivers, 25% had more than 18% of their area providing drift-feeding habitat for adult brown trout (Figure 14.10), and are examples of rivers where naturally occurring flows provide excellent trout habitat. On the other hand, 25% of the rivers had less than 7.5% of their area as trout feeding habitat and are examples of poor quality rivers when it comes to trout habitat.

Proposals to modify river flows can be assessed against this background – showing both the change in habitat quality and its value relative to other New Zealand rivers. This technique could be applied to benthic invertebrates as well as trout. Invertebrates are a major food source for many fish species and have been themselves used as indicators of stream health.

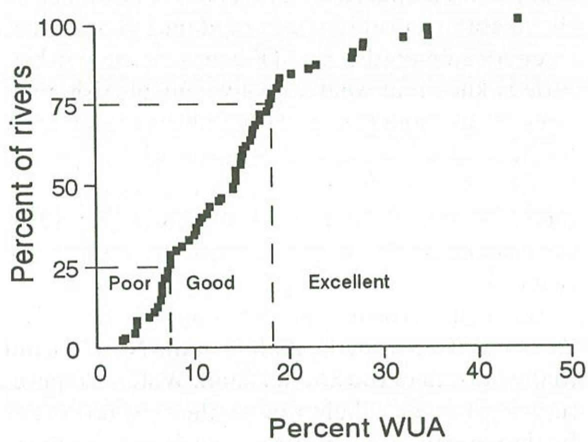

Figure 14.10 Comparison of the quality of drift-feeding habitat for large brown trout in 65 New Zealand rivers .

Habitat analysis could also be used to indicate the likelihood of changes in the composition of aquatic communities. One of the results of the "100 rivers" study was a model which predicted the probability of species composition and abundance of brown and rainbow trout from flow and other environmental factors (Jowett, 1990).

In summary, these results suggest that the primary factors to consider when predicting flow requirements are food and space, and in many situations the maintenance of suitable "food" and "space" habitat will maintain instream flow conditions for optimal production of a given species. However, instream habitat may not be the only environmental variable which changes if river flow changes, or if its catchment is modified. Water temperature, periphyton growth, and other aspects of water quality may change. The amount of suspended sediment and bedload may change, affecting the substrate composition and water clarity with resultant effects on the trophic status or food chain of the river. Channel "improvements" aimed at reducing flood levels and discharging flood waters often result in wider, shallower channels. Such channels provide poorer habitat quality and little shelter for trout both in normal and flood conditions. All of these effects can be modelled or at least estimated.

Acknowledgement

I would like to thank Gordon Glova, Eric Graynoth, John Hayes, Stephen Thompson, and Barry Biggs for their helpful comments and contributions to this chapter.

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15
Land-Use Impacts

B D Fahey and L K Rowe

Introduction

When the first Polynesian migrants came to New Zealand about a thousand years ago, forest covered 75% of the 26.5 million hectares of land. Forest clearance by fire and its replacement by grass and scrub during the Polynesian era meant that only 11.3 million hectares of forest remained by the time European settlers arrived (Cameron, 1962). From the early 1800s, tracts of lowland and hill country forests were cleared for pastoral farming, and by the 1950s the area of native forest had been reduced to 5.7 million hectares (about 23% of the original forest), much of it on steep, mountainous terrain (Thomson et al., 1972). About the same amount of land remains in tussock, although much of this has been severely depleted by burning, by grazing stock, and by infestations of noxious weeds and animals.

The planting of exotic forests over large areas began in the 1920s, and 1.3 million hectares had been established by 1990. Some of the land originally cleared for pastoral agriculture that had reverted to a mixture of native scrub and introduced species, notably gorse and broom, has been re-cleared and planted in pines. About 9 million hectares are currently in improved pasture. Other important changes in land-use include the draining and reclamation of wetlands (Stephenson, 1983) and urbanisation (McConchie, Chapter 18, this volume).

The country's economic growth has been dependent on the conversion of native forest and grassland to pasture and plantations, but in the process, rates of erosion have increased in many areas, stream channels have been modified, and flooding increased. Graphic accounts of the impacts of the large-scale deforestation of the New Zealand landscape and of the developing awareness of the need for soil and water conservation are given by McCaskill (1973) and Poole (1983). In 1988, the devastation caused by Cyclone Bola on the East Cape hill country (Figure 15.1) and floodplains emphasised the erosional consequences of forest removal.

In the 1950s there was a growing awareness that water, a multi-purpose resource, was also a finite commodity. Although it was clear that the large-scale conversion of one vegetation type to another could influence streamflow, little was known about the magnitude of these changes in New Zealand. Accurate and reliable scientific information on relationships between land-use and water yield are needed for sound management decisions on the allocation of scarce water resources and for evaluating the likely impacts of proposed land developments. This chapter briefly reviews some of the programmes set up to provide information on the hydrological impacts of land-use changes and of land-management practices. The effects of land-use change on water quality are discussed by Hoare and Rowe (Chapter 12, this volume), and have also been reviewed by McColl and Hughes (1981).

Figure 15.1 East Cape after Cyclone Bola (1988) showing the influence of vegetation type on landslide intensity. (Photograph by D Blake)

Some Fundamental Concepts

The Water Balance

The water balance is one way to depict the hydrologic cycle and, in its simplest form, is usually written as:

$$P = Q + E + G \pm S$$

where P is precipitation, Q is runoff (= total streamflow, also called water yield), E is evaporation (including transpiration), G is groundwater loss, and S is the change in soil-water storage (Figure 15.2). The equation can be re-arranged in the following form to highlight the impact of land-use change on runoff:

$$Q = P - (E + G) \pm S$$

Over a year, changes in runoff as a result of an alteration in land use will most likely reflect changes in the evaporation component of the water balance, as changes in soil-moisture storage and groundwater losses are likely to be small.

Evaporation from Vegetation

Evaporation losses are made up of three components: (a) evaporation of intercepted water from vegetation surfaces during and immediately after rainfall (wet-canopy evaporation, I), (b) transpiration of water from the soil through leaf surfaces

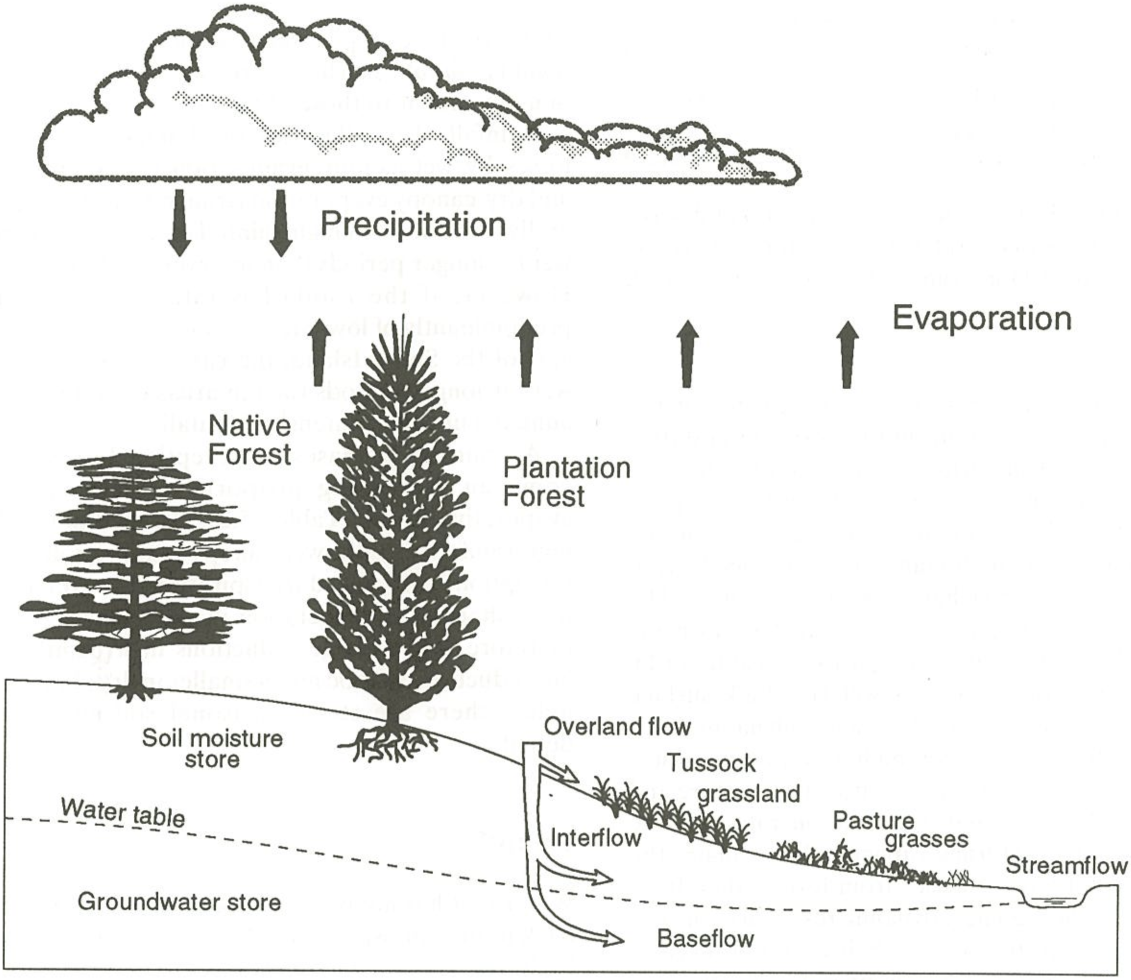

Figure 15.2 Components of the hydrological cycle and sources of streamflow

(dry-canopy evaporation, T), and (c) evaporation directly from the ground. Because the soil is normally shaded by plant canopies in forests and grasslands, interception and transpiration are responsible for most of the evaporative losses from these types of vegetation (Pearce, 1980).

The replacement of one vegetation type by another affects the water balance by altering both interception and transpiration. However, changes in catchment runoff tend to be dominated by differences in the amounts of rainfall interception and wet-canopy evaporation rather than by transpiration (Jackson, 1973; Calder and Newson, 1979; Pearce and Rowe, 1979; Pearce, 1980).

The most important factors controlling evaporation rates from vegetation are bulk surface resistance, r_c, and aerodynamic resistance, r_a (Monteith and Unsworth, 1990).

Bulk surface resistance represents the resistance water vapour must overcome in moving from the leaf interior to the surface via the leaf stomata. It is the reciprocal of the bulk canopy conductance, g_c as defined by Kelliher and Scotter (Chapter 8, this volume). When the leaf surface is dry the bulk surface resistance for a forest can be three times that for pasture; when wet it approaches zero (Table 15.1). Because of the high bulk surface resistance of dry forest and tussock grassland, their transpiration losses are normally lower than those of pasture and crops.

Resistance	Forest	Tussock	Pasture
r_c	75-150	160	50
r_a	5.5-7.5	7	30

Table 15.1 Bulk surface resistance (r_c) and aerodynamic resistance (r_a), in s m^{-1} for forest, tussock, and pasture (data from Table 2 in Murray et al. (1991)).

Aerodynamic resistance is a measure of the resistance water vapour must overcome in diffusing from the leaf surface to the atmosphere. This is a function of vegetation height and wind speed and is broadly equivalent to the reciprocal of the boundary-layer conductance, g_b as defined by Kelliher and Scotter (Chapter 8, this volume). The aerodynamic resistance for a forest is normally much lower than that for pasture (Table 15.1). Thus, when the canopy is wet (i.e. bulk surface resistance is near zero), the low aerodynamic resistance of forests promotes higher evaporation rates than those for short vegetation such as pasture and crops. It is this high evaporation rate, not the greater canopy storage capacity, that explains the higher interception losses from forests than from pasture. The high aerodynamic resistance for pasture and crops means that their evaporation rates are similar whether the canopy is wet or dry.

Interception and evaporation rates for tussock grasslands are similar to those from forests (Campbell and Murray, 1990) because of their similar aerodynamic resistance (Table 15.1, and Murray et al. (1991)). Tall, scrubby vegetation with "rough" canopies such as gorse, broom, and manuka, should also have aerodynamic resistances that are similar to those of forests.

Rainfall also influences the balance between losses by wet-canopy evaporation (interception) and dry-canopy evaporation (transpiration). Generally, vegetation in high-rainfall areas will remain wet for longer periods than in lower rainfall areas. However, if the rainfall is intermittent, and predominantly of low intensity, as in the south and east of the South Island, the canopy may remain wet for longer periods than in areas of equivalent annual, but higher intensity, rainfall.

As rainfall increases, interception losses become an increasing proportion of the total evaporative losses (Table 15.2). In medium and high-rainfall climates wet-canopy evaporation (interception) will exceed transpiration. Conversion from short to tall vegetation in wet climates can therefore lead to large reductions in streamflow, but reductions may be much smaller in dry climates unless there are strong seasonal soil-moisture deficits.

Runoff

Water that has not been evaporated or transpired back to the atmosphere is either stored temporarily in the ground or appears as streamflow. In water balance calculations (e.g. Table 15.2), runoff is expressed, like rainfall, in millimetres depth over the whole catchment, or as m^3 when considering catchment water supplies. Stream discharge, the volume of water passing a given point in unit time

Site	Water yield	Precipitation	Interception	Transpiration	I+T	Groundwater
Maimai	1550	2650	650	350	1000	100
Big Bush	900	1800	400	400	800	100
Purukohukohu	340	1485	385	660	1045	100
Taita	225	1130	350	430-630		?

Table 15.2 The water balance (mm) of some New Zealand native forests. Data are from Jackson (1973), Pearce and Rowe (1979), Dons (1987); underlined values were measured, the others were estimated.

(the product of water velocity and the stream cross-sectional area) is normally expressed in l s^{-1} or m^3 s^{-1}, and when plotted against time forms a stream hydrograph (Figure 15.3). Specific discharge is the stream discharge per unit area (e.g., l s^{-1} ha^{-1}), which is useful for comparing streamflow from catchments of different sizes.

Streamflow comes from precipitation into channels, overland flow (rainfall that flows over the ground surface before entering the stream channel), and subsurface sources (interflow through the soil and baseflow from ground water) (Figure 15.2). Streamflow from a river catchment can be rapid during storms (flood flow, stormflow or quickflow), but much water may be stored in the ground and released gradually between storms (low flow, baseflow or delayed flow).

Hewlett and Hibbert (1967) proposed a simple procedure for partitioning streamflow into quickflow which leaves the catchment rapidly during and immediately after storms, and delayed flow occurring between storms (Figure 15.3). This procedure has been widely used overseas (e.g., Harr et al., 1975; Hewlett et al., 1977; Wright et al., 1990) and in New Zealand (e.g., Pearce and McKerchar, 1979; Taylor and Pearce, 1982; Jackson, 1987; Smith, 1987). The proportion of total streamflow appearing as quickflow in New Zealand catchments ranges from near zero in areas of porous pumice soils where most rainfall is quickly absorbed, to over 70% in steepland catchments where shallow soils overlie less permeable bedrock (Pearce and McKerchar, 1979).

Assessing the Impact of Land-use Change

Catchment Experiments

The traditional way of monitoring land-use change is by catchment experiments, in which rainfall and stream discharge are monitored in selected catchments for specific periods around a planned altera-

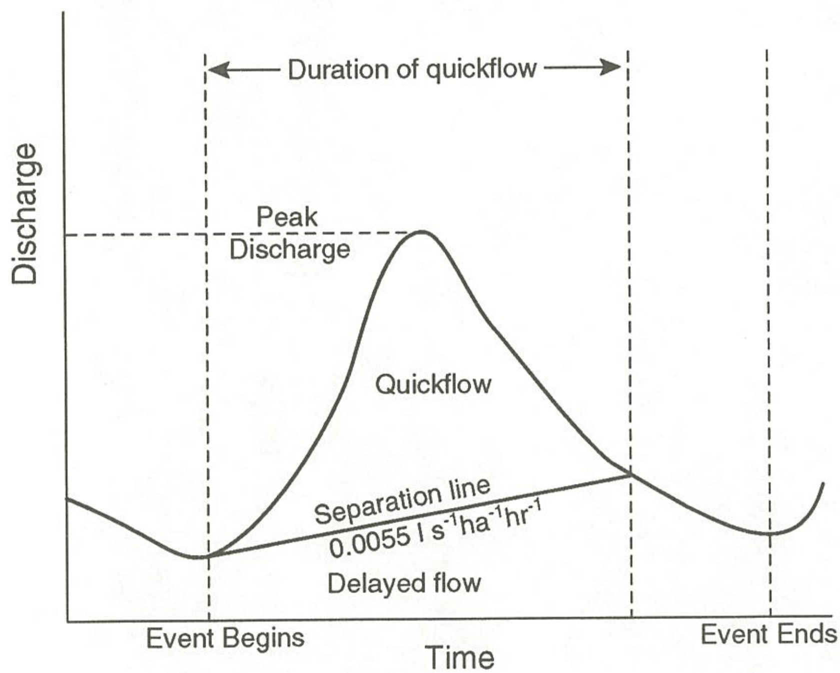

Figure 15.3 A stream hydrograph showing the hydrograph separation method of Hewlett and Hibbert (1967)

tion in land use. One of three basic approaches is normally adopted.

The simplest is the "before and after" approach, in which components of the water balance of a catchment are measured until its behaviour can be readily predicted. The catchment is then altered and subsequent measurements are compared with the predicted stream responses if the catchment had remained unaltered.

In the "paired-catchment" approach, rainfall and stream discharge are measured in two neighbouring catchments of similar size and shape. Once they are calibrated, one of the catchments is "treated" (i.e., subjected to a particular land-use change or management option), and the impacts of the treatment are measured as the departure from the predicted behaviour determined by comparison with the control or untreated catchment. When more than two catchments are used in the study it becomes a "multiple-catchment" experiment (Figure 15.4).

Sometimes the water balances of neighbouring catchments with different land uses are compared directly, without a "before and after" sequence or an initial calibration period. Relative differences can be identified but absolute differences can not.

Figure 15.4 The Maimai experimental catchments near Reefton: an example of a multiple-catchment study. (Photograph by H Hemming)

Catchment Programmes for Water-Resource Evaluation

Experimental catchments were established at Moutere, Makara, and Taita (Figure 15.5) in the 1950s to examine the effects of land-use change and management practices on the hydrology of the catchment. Other catchment studies at Puketurua, Purukohukohu, and Otutira followed under the UNESCO-sponsored International Hydrological Decade (IHD) programme in 1965.

The consequences of pasture management were studied at Makara (Toebes et al., 1968; NZNCIHD, 1969; Yates, 1971) and Moutere (Anon, 1973; NZNCIHD, 1969), land-use change at Moutere (Scarf, 1970; Duncan, 1980; Waugh, 1980), and different land-uses at Puketurua (Waugh, 1980; Waugh et al., 1981), Purukohukohu (Jackson, 1973; Waugh 1980; Dons, 1987), Otutira (Selby, 1972), Taita (Jackson, 1973; Claridge, 1980), and Moutere (Duncan, 1980; Waugh, 1980).

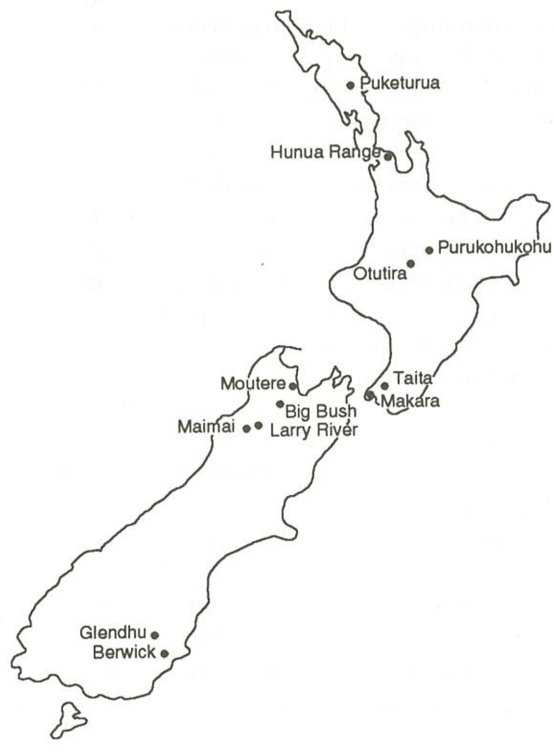

Figure 15.5 Location of hydrological experimental catchment studies in New Zealand

In the early 1960s, the Representative Basin network was established to provide benchmark water-resource data for each of the 90 hydrological regions of New Zealand (Toebes and Palmer, 1969). In 1978, the New Zealand Ministry of Works (MOW) launched the Land-use Basin Programme, which paired catchments for direct comparison of hydrological responses to different land uses in different hydrological regions (Waugh, 1981; Davoren, 1985; 1986).

The Land-use Impacts section of the Forest Research Institute (now Landcare Research New Zealand Ltd) operates three experimental catchment studies in the South Island. The hydrological consequences of harvesting beech-podocarp-hardwood forest on steep hill country and establishing *Pinus radiata* plantations have been investigated at Maimai since 1974 (Pearce et al., 1976, 1980; Rowe and Fahey, 1991), and at Big Bush since 1979 (Pearce et al., 1982) (Figure 15.5). At Glendhu in east Otago, a paired-catchment study has been made of the hydrological impacts of converting mid-altitude tussock grassland to pine plantation (Pearce et al., 1984; Fahey and Watson, 1991).

Simulation Models

Experimental catchments have also been used for process-based studies on soil-water-plant relationships. Investigations of evaporation, interception, and streamflow generation help explain why catchments respond in a particular fashion to a given change in vegetation cover, and can increase the confidence with which results from a given catchment can be extrapolated to other areas.

Catchment and process-based investigations have fostered the development of hydrological models, which simulate the water balance of the catchment, starting with the initial input from rainfall and its subsequent partitioning into streamflow, evaporation, and storage.

Traditionally, hydrological models have relied on our understanding of the underlying physical processes governing streamflow generation, for example, the Stanford Model (Crawford and Linsley, 1966). Substantial progress has been made in modelling evaporation, one of the key components of the water balance, and in the development of biophysical models for simulating hydrological processes within a catchment (e.g., Whitehead and Kelliher, 1991).

If the results from hydrological models match measured water balance components derived from catchment studies, the models should be useful in predicting the effects of land-use change, and in management where catchment data are not available. In addition, they should allow the consequences of different land-management options to be explored.

Major Land-Use Conversions

Changes in land use have occurred in the past and are still occurring; for example, reforestation of the East Cape hill country, pasture improvement in tussock grasslands, reversion of pasture to scrublands, and conversion of scrublands to pine forests. These changes, especially in small catchments, affect the hydrological regime.

Native Forest to Pasture

Because evaporation rates from wet forest canopies are higher than those from wet pasture, and forests can extract moisture for transpiration from a greater depth of soil, forested areas should yield less water than pasture areas. Studies in New Zealand tend to confirm this.

Pasture catchments have higher annual water yields, peak discharges, and quickflows than comparable catchments in native or pine forest (Davoren, 1986). However, at low stream discharges pasture may have smaller flows than native forest but higher flows than pine forests.

At the Purukohukohu Experimental Basin for example, annual water yield from pasture was 200 mm (60%) higher than that from native forest (Table 15.3) which reflects the lower evaporation loss from pasture.

Quickflow from the pasture catchment at Purukohukohu was almost 10 times higher than from native forest, and delayed flow was less sustained and lower overall than from forest.

The differing streamflows from pasture and forest are not always due solely to differences in vegetation cover. For example, one reason the pasture catchment at Purukohukohu yields up to 10 times more quickflow than the one in native forest is the lower permeability of the topsoil caused by stock trampling (Jackson, 1973).

Tussock Grasslands to Pasture

Because the evaporative characteristics of tussock grasslands and native forests are similar, we would expect water yields to increase by similar amounts after the conversion of one or the other to pasture. However, native tussock has lower transpiration rates than improved pasture. Replacing a tussock cover with improved pasture may increase transpiration, especially in summer, which in turn may reduce late summer streamflows.

There is some support, but no published data, for this interpretation. South Canterbury and East Otago, despite differences in rock type and topography, have higher annual water yields from tussock catchments, even in a depleted condition, than from their counterparts in pasture (Davoren, 1986). The biggest difference occurs during low streamflow, with tussock catchments having con-

	Pasture		Native forest	
Precipitation	1427		1484	
Quickflow	74	(5)	8	(1)
Delayed flow	469	(33)	331	(22)
Total streamflow	543	(38)	339	(22)
Evaporation	784	(55)	1045	(70)
Groundwater loss	100	(7)	100	(7)

Table 15.3 Annual water balance (mm) for two catchments at Purukohukohu, one in pasture and the other in native forest. Percent of precipitation in brackets. Data from Dons (1987).

sistently higher and more sustained levels of flow. By contrast, peak discharges differ little between pasture catchments and tussock catchments in good condition.

Native Forest to Pine Plantation

Depending on stocking densities, the interception storage capacity and aerodynamic resistance of a mature pine stand should not be much different from that of a native beech or podocarp stand. Thus, the replacement of native forest with plantation species should not alter the catchment water balance in the long term. However, in the short term, water yield should increase dramatically immediately after the removal of the original forest cover.

Short-term changes: The immediate impact of land clearance, whether scrub or forest, before establishment of a new vegetation cover is an increase in streamflow as evaporative losses (interception from wet canopies or transpiration from dry canopies) are greatly reduced. This response appears as an increase in quickflow, as higher peak discharges for smaller storms and as increased delayed flow. In New Zealand, annual water yield can increase up to 650 mm in the first year after clearance (Table 15.4).

Increases in water yield after harvesting beech-podocarp forest in north Westland (Pearce et al., 1980; Rowe and Fahey, 1991) and south-east Nelson (Pearce et al., 1982) are large compared with many of those observed overseas (e.g., Bosch and Hewlett, 1982). They are, however, consistent with the high rainfall regime (average rainfall 2500 mm yr^{-1}) and associated interception losses (up to 750 mm in a wet year) recorded under undisturbed forest at Maimai (Rowe, 1979). At Big Bush, with an average rainfall of 1530 mm yr^{-1}, up to 535 mm yr^{-1} of water previously lost by interception was

	Area cleared (%)	Precipitation (mm)	Water yield increase (mm)	Reference
Manuka				
Puketurua	100	1565	172	1
Hunua Range	100	1880	179	2
Gorse				
Moutere 8	100	1069	168-180	3
Moutere 13	100	1069	283-302	3
Native Forest				
Maimai 5	100	2625	550	4
Maimai 8	95	2827	260	4
Maimai 13	90	2625	200	4
Maimai 7	100	<u>1930</u>	650	5
Maimai 9	80	<u>1930</u>	540	5
Big Bush DC1	83	1305	373	6
Big Bush DC3	74	1305	331	6
Big Bush DC4	94	1305	420	6

Table 15.4 Increases in first-year water yield as a result of land clearance
Note: Data underlined were pro-rated from 19-month totals
References: 1 Waugh (1980); 2 Herald (1979); 3 Duncan (1980); 4 Rowe & Fahey (1991); 5 Pearce et al. (1980); 6 Pearce et al. (1982)

available for additional water yield after the forest was harvested (Rowe, 1983). At Maimai, most of the increased water yield from harvested catchments came in the form of extra quickflow during small storms. Harvested catchments yielded up to 3 times more quickflow per storm than the control catchment and had peak discharges up to 67 % higher. For large storms, the increase in quickflow was only 15-20% and in peak discharges about 30 % (Pearce et al., 1980). At Big Bush, however, much of the increased streamflow after harvesting came from increased delayed flow.

Long-term changes: Regrowth of vegetation (grasses or invading scrub species and the young tree crop) causes catchment water yield to progressively decline as interception and transpiration increase. Initially, the decline is related to invasion of the catchment by grasses, gorse, broom, bracken and other weeds, rather than to growth of the planted tree seedlings, which will be too small for a number of years to affect water yield. Even after canopy closure, evaporation losses from the understorey can be high, up to 94 mm yr^{-1} for a 13 year-old *P. radiata* stand at Purukohukohu (Whitehead and Kelliher, 1991). The decline in water yield is variable for each region, depending on biomass production rates and the precipitation regime (Table 15.5).

The conversion of native scrub to pine forest in the Hunua Ranges south-east of Auckland caused an initial increase in water yield of 157 mm yr^{-1}. The yield then decreased at a rate of about 70 mm yr^{-1} until after 7 years of forest growth it was 70% of that before scrub clearance (Herald, 1979). At Maimai, water yields returned to pre-treatment levels after 3-6 years of vegetation regrowth (mainly bracken, Himalayan honeysuckle and planted pines) (Figure 15.6). Except for one catchment, water yields continued to decline below pre-treatment levels, eventually stabilising at about 200 mm yr^{-1} less than estimated for native forest. At this stage of regrowth, water use is dominated by that of the scrub species.

Locality	Water yield decrease (mm)	Time (years)	Reference
Hunua Range	480	7	Herald (1979)
Moutere 8	290-300	8	Duncan (1980)
Moutere 13	280-300	5	Duncan (1980)
Maimai M5	730	7	Rowe & Fahey (1991)
Maimai M8	260	7	Rowe & Fahey (1991)
Maimai M13	200	2	Rowe & Fahey (1991)

Table 15.5 Long-term water yield decrease after afforestation with pines following land clearance.

Water balance calculations for the Purukohukohu catchments showed water yield from an 8 to 10-year-old *P. radiata* stand to be similar to that for an adjacent native forest (18 % and 22 % of rainfall, respectively) (Table 15.6). Quickflow from the pine catchment was higher (2.0 % of rainfall compared with 1.0%) but was low in

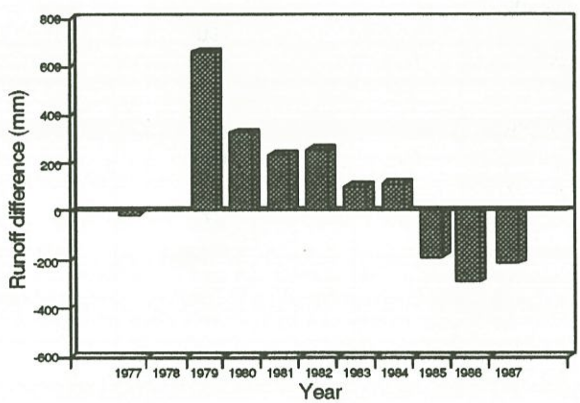

Figure 15.6 Difference in annual water yield between Maimai catchments M6 (left in native forest) and M5 (clearfelled, hauler-logged, desiccant sprayed and planted in 1978).

	Native forest		8-10-year old pine	
Precipitation	1484		1398	
Quickflow	8	(1)	31	(2)
Delayed flow	331	(22)	223	(16)
Total streamflow	339	(22)	254	(18)
Evaporation	1045	(70)	1054	(75)
Groundwater loss	100	(7)	100	(7)

Table 15.6 Annual water balance (mm) for two catchments at Purukohukohu, one in native forest and the other in pines. Percent of precipitation in brackets. Data from Dons (1987).

both catchments compared to other New Zealand catchments, which can be up to 40 % (Pearce and McKerchar, 1979). The low values at Purukohukohu reflect the high permeability of the ash and pumice-derived soils. Delayed flows were sustained longer at higher levels under the native forest than under the pine forest.

Annual water yields from native and exotic forest catchments at Nelson, Wellington and the Central North Island did not differ significantly (Davoren, 1986). However, quickflow from native forest catchments was lower, and low flows were more persistent than from their counterparts in pines (Figure 15.7).

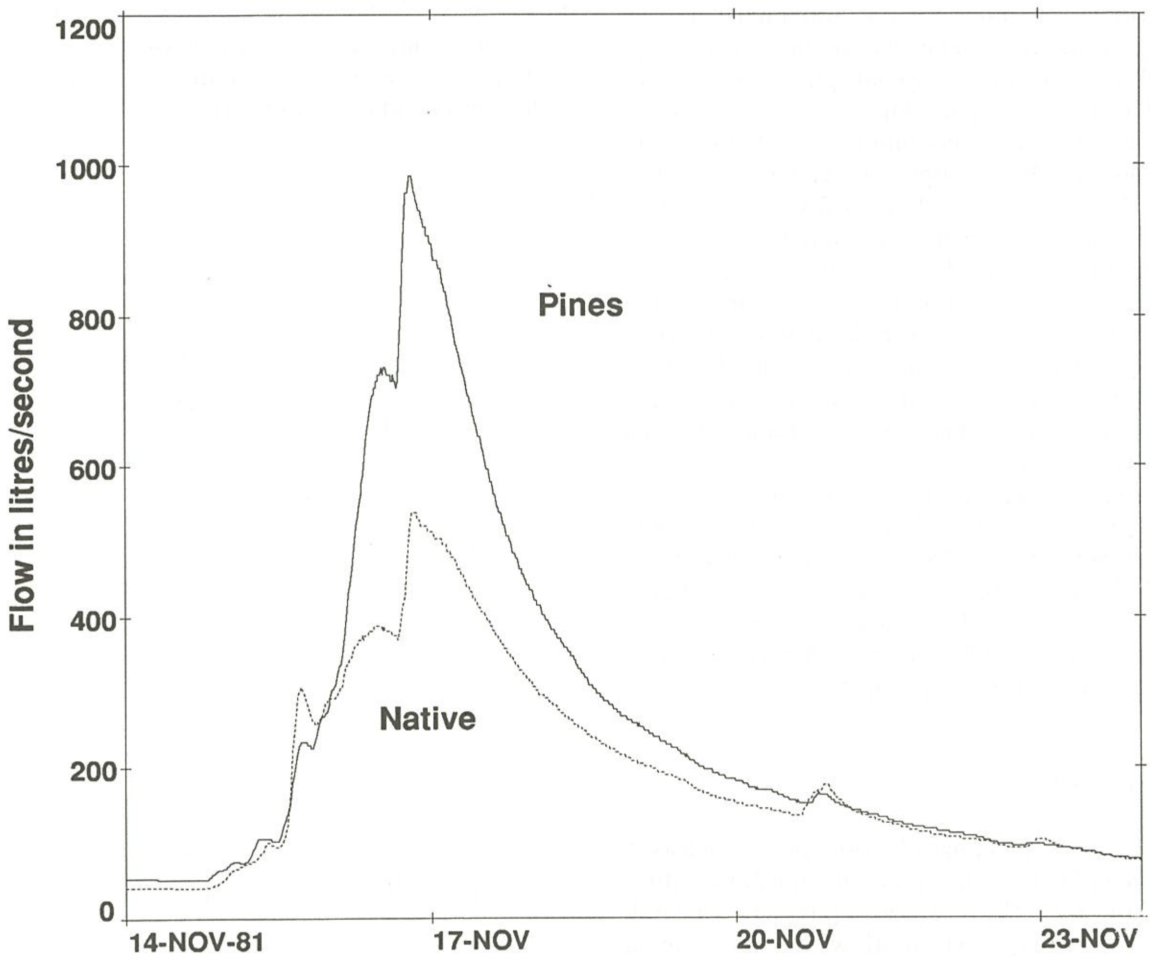

Figure 15.7 Typical storm hydrographs for catchments in pine forest (Graham Creek, 474 ha) and in native forest (Hunters Gully, 502 ha) at Nelson. Note the lower quickflow from the native forest catchment.

Minor Land-Use Conversions

Tussock Grassland to Pines

Because tall tussock grasslands and forests have similar aerodynamic resistances, their wet-canopy evaporation rates may also be similar. Thus, in theory, the conversion of tussock to pine plantation should have little impact on water yields.

However, it is unlikely that the tussock cover will be completely, or even partially, removed before planting. The growth of pines in an area of tussock grassland thus adds another tier to the canopy cover that will also intercept and transpire water. The understorey of tussock will eventually die, but will continue to intercept some water not caught by the growing trees. Thus we would expect water yields from tussock grasslands planted in pines to decline as the trees develop.

Data from two medium-sized catchments in upland east Otago support this conclusion (Fahey and Watson, 1991). After a 3-year calibration period one catchment was planted in *P. radiata*. Little difference in the streamflow regime of the two catchments was observed for 6 years. In the eighth year (1989), however, the annual water yield for the catchment in pines was 130 mm less than that for the catchment left in tussock, representing a 19% reduction. This trend continued through 1990 and 1991 (Figure 15.8).

Peak discharges in small storms were also influenced by tree growth, showing an average reduction of 50% (Figure 15.9). For storms with large peak discharges the reduction was 22%. Quickflows were 25% and 32% lower on average in the planted catchment for small (< 20 mm) and large (> 20 mm) storms, respectively.

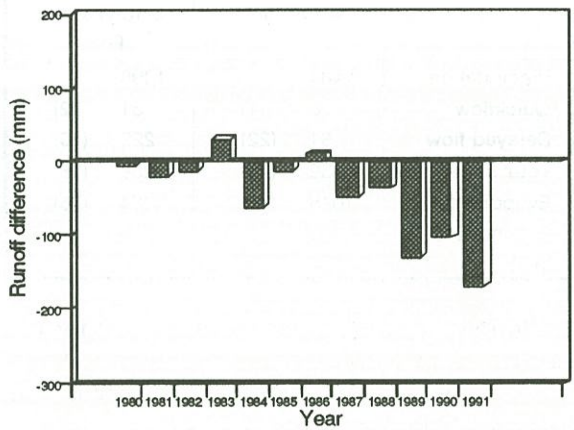

Figure 15.8 Difference in annual water yield between the control catchment in tussock grassland and the catchment planted during 1982 in *P. radiata* at Glendhu, east Otago.

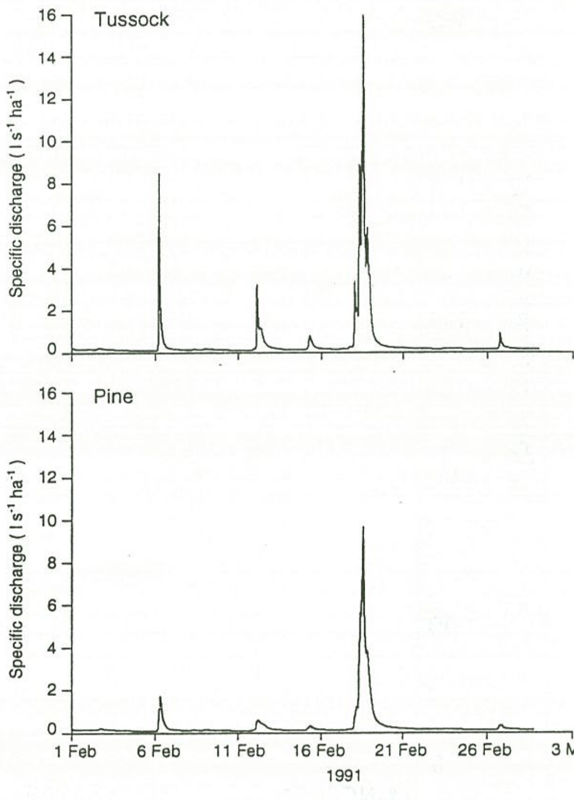

Figure 15.9 Storm hydrographs from catchments in tussock grassland and in 10-year-old *P. radiata*, Glendhu, east Otago.

Pasture to Pines

Any extensive plantings of exotic species such as *P. radiata* or Douglas fir in areas previously in pasture should, like their counterparts in tussock grasslands, cause streamflows to decrease in response to higher evaporation rates from forest canopies.

Annual pasture water yield at Purukohukohu was 289 mm higher than from a neighbouring catchment in pines; evaporation was 270 mm lower (Table 15.7). At Berwick in east Otago, total annual water yield from pasture was 231 mm more than from 14-year-old pines.

Quickflow at both Purukohukohu and Berwick was higher from the pasture catchment. For example, at Berwick, quickflow averaged 30% of streamflow from pasture, whereas for pines it was only 13% (Smith, 1987). Stream discharge was sustained at a higher level in a pasture catchment, being greater than 3.2 $l\ s^{-1}\ km^{-2}$ for 80 % of the time whereas the corresponding level for the planted catchment was about 1.5 $l\ s^{-1}\ km^{-2}$.

Scrub to Pines

Tall scrub and forest covers have similar aerodynamic resistances. Thus, apart from increases in the first few years after scrub clearance (Table 15.4), water yields should not be much different than those before conversion. However, streamflows from planted areas previously in scrub will depend on stocking densities and pruning and thinning practices.

The water yield from a catchment at Moutere initially doubled in response to gorse clearance, but had almost returned to pre-treatment levels within 4 years of planting pines (Figure 15.10). The decline in yield continued for a further 4 years (at which time the canopy had closed) when it was 62% lower than it would have been if the catchment were still in gorse (Duncan, 1980).

Scrub to Pasture

After conversion of manuka scrub to pasture at an experimental catchment at Puketurua, the annual water yield increased by about 172 mm in the year after clearance and to 255 mm in the second year (Figure 15.11). Similar results were obtained after kanuka scrub removal in the Hunua Ranges, (Barton, 1972; Herald, 1979) and gorse removal at Moutere (Duncan, 1980).

At Moutere, most of the increase came from small- and medium-size storms (Scarf, 1970). After conversion, peak discharges were higher in the recently cultivated catchment for small- and medium-sized storms; for a 3 $l\ s^{-1}\ ha^{-1}$ peak in one control catchment, the equivalent at the treated catchment increased from about 0.6 $l\ s^{-1}\ ha^{-1}$ to about 5.4 $l\ s^{-1}\ ha^{-1}$. For flood peaks above about 45 $l\ s^{-1}\ ha^{-1}$, there was little change, indicating that the removal of scrub had little influence on stream discharge in severe storms.

Purukohukohu	Pasture		Pine (8-10 years)	
Precipitation	1427		1398	
Stormflow	74	(5)	31	(2)
Delayed flow	469	(33)	223	(16)
Total streamflow	543	(38)	254	(18)
Evaporation	784	(55)	1054	(75)
Groundwater loss	100	(7)	100	(7)
Berwick Forest	Pasture (G1)		Pine (F2) (14-21 years)	
Precipitation	978		1040	
Total streamflow	396	(40)	165	(16)
Evaporation	582	(60)	875	(84)
Streamflow (6 yr total)	2182		973	
Stormflow (6 yr total)	601		139	
Stormflow/stream flow ratio	28		14	

Table 15.7 Annual water balance (mm) for pasture and pine forest catchments. Percent of precipitation in brackets. Data from Dons (1987) and Smith (1987).

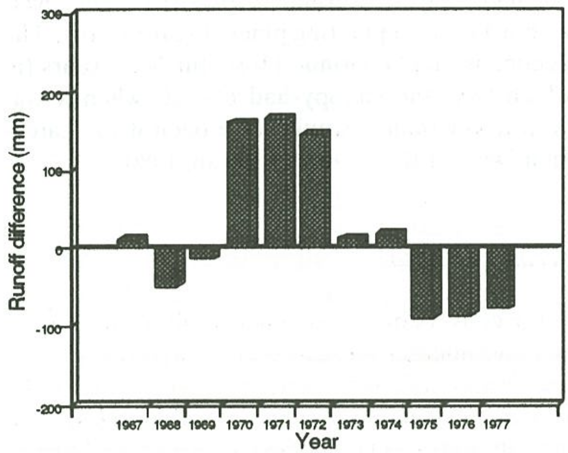

Figure 15.10 Departure of annual water yield from that predicted for Moutere catchment 8 after a calibration period in gorse, followed by clearance in 1970 and planting in pines. Data from Duncan (1980).

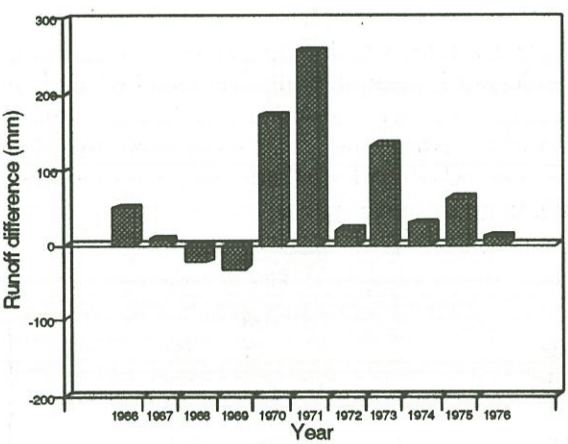

Figure 15.11 Departure of annual water yield from that predicted for one of the Puketurua catchments after scrub clearance in 1970, cultivation, and pasture growth. Data from Waugh (1980)

Pasture to Scrub

After initial clearance of forest for farming, some areas of New Zealand have developed into scrublands with either native (e.g., manuka, kanuka, bracken) or introduced (gorse and broom) scrub species. Because of similarities in aerodynamic resistance between scrub and forest covers, the hydrological differences between scrub and pasture or crop catchments should be similar to those between forest and pasture.

At Moutere, streamflow trends were similar to those for forest to pasture conversions. Streamflow from gorse and stocked pasture catchments was 14 % and 30 % of rainfall (1100 mm), respectively (Scarf, 1970). Peak discharges from gorse catchments averaged 78 % lower than from pasture catchments (Duncan, 1980). In contrast, in a "before and after" experiment at Puketurua, annual water yield from a catchment that had been in pasture for 4 years was similar to the yield when the catchment had been in manuka scrub, about 660 mm (44 %) of 1500 mm rainfall (Waugh, 1980).

Pasture Improvement and Grazing

Many areas in New Zealand have been subjected to management changes rather than changes in vegetation type. These changes include pasture improvement, soil conservation measures, and differing grazing regimes, and featured in many of the early Experimental Basin research programmes.

At Makara, for example, oversowing and topdressing of two of four intensively grazed catchments caused a reduction in average annual water yield over the 5-year evaluation period of 40 - 50 % and a reduction in peak discharge of 75 % (Toebes et al., 1968). Most of this reduction occurred during the smaller storms (peak discharges <30 mm), but larger storms also showed lower peak discharges. Not only was flow reduced, but fewer streamflow events were recorded in the improved catchment than in the control. Yates (1971) attributed these changes to increased interception, infiltration, and transpiration.

At Otutira, a runoff-plot study has shown that grazed pastures tend to have higher surface runoff than comparable ungrazed pasture (4.6 % and 0.7 % of precipitation, respectively). Soil compaction by stock and machinery lowered soil porosity and thereby decreased rainfall infiltration (Selby, 1972).

Wetland Development

Wetlands throughout the world are under constant threat of development. In New Zealand, wetlands have provided peat and sphagnum moss for horticulture, flax, waxes, kauri gum, ironsands and coal for industry, and, after drainage, land suitable for agriculture and forestry. Less than 10 % of New Zealand's former wetlands remain undeveloped, and many of these are under pressure (Stephenson, 1983).

Wetlands have considerable hydrological value. The large mountain wetlands in eastern Otago help sustain catchment streamflow. Peatlands such as the Whangamarino Swamp in the Waikato and the Kopuatai Peat Dome on the Hauraki Plains are essential for flood control, storing flood waters and gradually releasing them to local rivers. Wetlands also can extract nutrients from river waters, thereby improving downstream water quality (Stephenson, 1983).

The pakihi wetlands on the West Coast of the South Island have been drained for farm development and establishment of pine plantations. Soils are often between 0.5 - 1 m deep, and subsoils are perennially saturated, with slow lateral or vertical movement of soil water. An undisturbed control catchment at the Forest Research Institute's Larry River study area, near Reefton, showed that these natural wetlands are highly responsive to rainfall. During 3 years of measurements (with annual rainfall averaging 2440 mm), the water yield averaged 1580 mm in over 200 floods, with quickflow accounting for over 70 % of the total (Jackson, 1987).

To prepare pakihi land for tree planting, a bulldozer with a V-shaped blade removes a 0.3-0.4 m strip of soil, which rolls outwards to form mounds 0.5-1.0 m high on each side of the strip. The tree seedlings are then planted on the mounds, which allows adequate drainage for root development. The drainage operation disturbs about 2/3 of the catchment area, with 1/3 in clean straight drainage channels, 1/3 in newly formed mounds, and the remaining 1/3 undisturbed (Figure 15.12).

Afforestation of pakihi can lead to increased water yields, and higher and more frequent peak flows. In the 18 months after drainage of a Larry River catchment, three times as many peak discharges greater than 10 l s^{-1} ha^{-1} were measured than in the undisturbed control catchment. In many larger events, peak flows from drained catchments were 2-4 times greater and quickflow were up to 30% greater than those from the undrained catchment, whereas in the pre-treatment period they were similar. Small rainfall events (< 10 mm) often produced streamflow from the drained catchments but not from the control catchment.

Large Catchments

For extrapolation of results from small experimental catchments to large catchments, other hydrologic factors must be taken into consideration. For example, floods in large catchments are influenced more by channel storage characteristics, which may mask the effects of high rainfall intensity, quickflow and land-use (Burton, undated; Chow, 1964).

Apart from urbanisation, large-scale afforestation is the one well-documented land-use change. Generally, total water yield tends to decrease as a catchment is planted, or increase with deforestation.

In the central North Island between 1964 and the early 1980s, over 250 km^2 (28%) of the 906 km^2 Tarawera River catchment was planted in pines. The previous cover had been light scrub (60%) and native forest (40%) (Dons, 1986). Simple flow models revealed that afforestation could account for 4.5 m^3 s^{-1} of the 13% decrease in mean discharge between 1964 and 1981 compared with the calibration period. The rest of the reduction (6.4 m^3 s^{-1}) was attributed to lower rainfall during the afforestation period.

This result is similar to those reported overseas. In 10 river catchments in the Southern Piedmont of the south-eastern United States, which range in area from 2820 to 19450 km^2 and in which 10 - 28% had been afforested, water yields declined by 30 - 100 mm (4 - 21%), but there was little correlation between the degree of afforestation and the decrease in yield (Trimble and Weirich, 1987). In the 200 km^2 Queens River catchment, Pretoria, South Africa, as the area in forest increased from 12 to 55% over a 23-year period, water yield

Figure 15.12 Land preparation for planting trees on West Coast Pakihi wetlands (Photograph by R J Jackson)

decreased at an average rate of about 4.3 mm yr^{-1} (Pitman, 1978).

Harvesting practices in headwater areas probably have little influence on downstream floods in major storms, as the volume of precipitation far outweighs the minor soil moisture and interception differences between cut and uncut areas (Rothacher, 1971; Brown, 1973; Harr et al., 1975). Duncan (1986) was unable to detect a change in the size of peak flows from the 232-km^2 Deschutes catchment in Washington State, which had been harvested over a 30-year period at an average rate of 344 ha yr^{-1}. The lack of change at this catchment scale may have reflected the continuing regeneration of the forest and the distribution of the harvesting operation throughout the catchment.

Simulation Models

Simulation models are used to predict the consequences of a land management practice on streamflow, and they are particularly useful where streamflow data are unavailable or unreliable. For example, from a numerical model of the energy balance, Fitzharris (1974) predicted that conversion of Central Otago tussock grassland to introduced pasture grasses would increase evaporation and therefore decrease streamflow. Another model, based on data collected at the Glendhu experimental catchments (Pearce et al., 1984) predicted that this conversion of tussock to introduced grasses may increase summer evaporation, which in turn may reduce low flows in late summer (Fahey et al., 1991).

When used to assess streamflow from Deep Stream in the Lammermoor Range, the source of about half of Dunedin's daily water supply, the model predicts that daily yield is sufficient in an average year to sustain the minimum required water supply. However, if the three summer months are drier than average, discharge at Deep Stream may fall below the minimum requirements. The model further shows that restoration of the tussock cover in the key water-producing areas would reduce the likelihood of water shortages.

Process-based models that have been calibrated against measured transpiration, evaporation, and soil drainage can be applied to a variety of land-management issues. For example, experimental catchment studies have shown that converting pasture to pines reduces water yields and subsequent clearfelling leads to more water in streams. It is less clear what happens to streamflow when trees are thinned and pruned during the growth phase of a crop rotation. The changes are unlikely to be as great, but they may still affect downstream users relying on water for irrigation or municipal supplies.

Whitehead and Kelliher (1991) combined the Penman-Monteith equation (Monteith and Unsworth, 1990) for estimating transpiration losses with the Rutter model of interception (Rutter, et al., 1971) to calculate the annual canopy water balance of a *P. radiata* stand near Rotorua before and after thinning. Simulation studies showed that a minor change in the timing of thinning and pruning a pine plantation could have increased water yield from 42 to 53% of annual rainfall.

Summary and Conclusions

Land-use changes can have a significant influence on the water balance of small catchments and the production of streamflow.

The removal of a forest or scrub cover and its replacement with pasture will increase annual water yields. The conversion of gorse or native scrub to pasture, for example, can cause streamflow to double, mostly through increased quickflow during small and medium storms. Once established, pasture improvement by oversowing and top dressing can reduce annual yields by up to 50%. Peaks discharges may fall by 75%.

The depletion of tussock grasslands by burning and grazing, or their replacement with introduced grasses, will reduce water yields primarily through the effect these changes have on low flows. Peak discharges may increase, but low flows are typically smaller and less reliable.

In medium-to-high rainfall areas, the replacement of native forest with pines can cause water yield to increase by 400-600 mm yr^{-1} (75%) in the first few years after clearfelling. Most of the increase comes from small and medium-size storms through higher quickflows. A return to pre-treatment levels is normally achieved after 5 years, and may eventually be lower than that before disturbance, perhaps by as much as 200 mm. However, quickflows may remain higher on average, whereas delayed flows are smaller.

A similar response occurs when areas in gorse and scrub are cleared and planted in pines. Water yields may increase by up to 100% immediately after clearance, but quickly decline after planting, eventually by as much as 60% after canopy closure.

When pines are planted in tussock grassland, annual yields will decrease by 200-300 mm (20-30%) within 6 - 8 years, mostly from reduced discharge in small storms. Further reductions can be expected as the trees mature but the final figure will depend on the thinning and pruning regime.

For pasture lands planted in pines annual yields may be reduced by 50% once the canopy closes. Low-flows may fall by as much as half, and peak discharges by 80%. Although these changes may be beneficial as a means of flood protection they may have serious consequences where downstream users are dependent on sustained streamflows for irrigation, hydro-electric power generation, and municipal supplies.

Acknowledgements

We would like to thank Rick Jackson, David Murray and Maurice Duncan for their helpful comments on earlier drafts of this chapter, Joanna Orwin for editorial assistance, and Thomas Pearson for assisting with the preparation of the figures.

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16
River Morphology

M Paul Mosley

Introduction

River channels comprise a very small proportion of the total land surface of New Zealand, but they are a very significant component of the landscape, and of the hydrological cycle. River channels and drainage networks are intimately related to streamflow (chapter 7) and sediment load (chapter 13), because on the one hand their shapes reflect the flows and sediments which they convey, and on the other they themselves modify the characteristics of streamflow, such as hydrograph shape.

Qualitative and descriptive approaches to explaining river channel characteristics are well-developed, based on the approaches used in historical geology and geomorphology (Schumm, 1972, 1977). In New Zealand, this approach is exemplified by the work of Cotton (1958) and Gage (1980). However, a strongly quantitative approach to fluvial geomorphology has also been established, based on the early morphometric work of Horton (1945) and Strahler (1952), the hydraulic geometry developed by the United States Geological Survey (summarised in Leopold et al, 1964), regime theory (an empirical rather than theoretical methodology of open channel design; Blench, 1957), and river mechanics and hydraulics.

Rivers in New Zealand are similar to those in other countries, when climatic and geological conditions are the same. For example, the wide braided rivers of the South Island look very much like braided rivers which flow from glaciers in Canada and the United States. Meandering rivers like the Manawatu and Waipaoa, which wind across low gradient, silty flood plains, are similar to European rivers like the Trent, in England. The most commonly seen type of river in New Zealand, with a gravel bed, riffles, and a moderately straight channel, is very similar to many in Switzerland or Scotland, which carry gravel bedload and experience periodic floods caused by heavy rainfall in the headwaters.

New Zealand is geologically young and active, mountainous, and has a humid temperature climate with frequent and abundant rainfall in most parts of the country. The resulting high rates of erosion, steep valleys, and frequent floods and freshes give rivers which are, in international terms, particularly active, and ever-changing.

Classification of Rivers

Rivers form a continuum of types, from tortuous meanders through to braided, which reflect sediment supply and character, the stability of flows, and channel gradient, (Figure 16.1; Mollard, 1973). Examples of all these types, from tortuous meandering through to braided, may be identified in New Zealand.

Several classification schemes for rivers have also been developed, which may be useful in clarifying the relationships between channel shape and the controlling variables (Schumm, 1977). Nevins (1965) classified New Zealand rivers on the basis of "the fundamental relationship between the

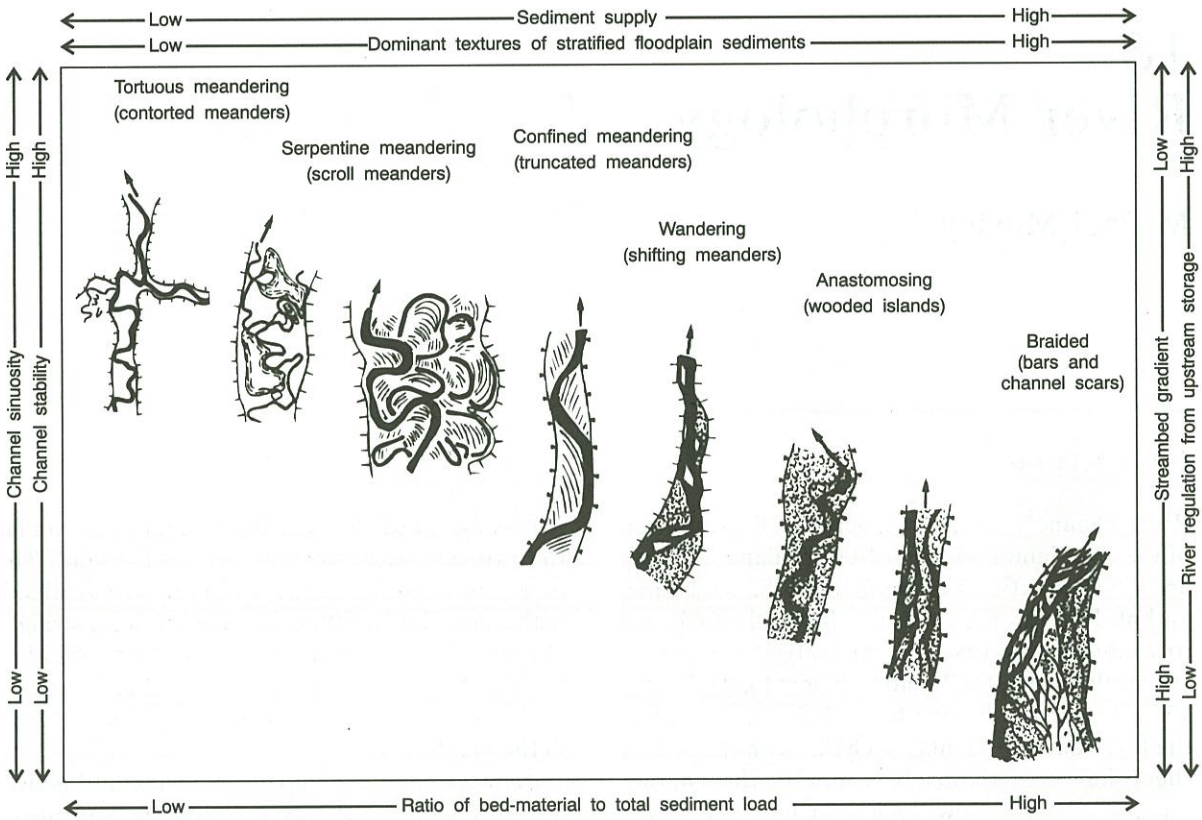


Figure 16.1 The continuum of river channel types, and controlling variables (after Mollard, 1973).

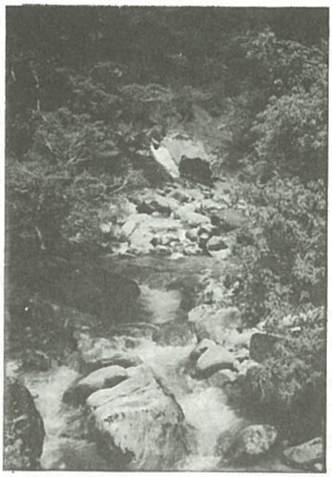
quantity of detritus (sediment) carried by rivers and its dominant grain size, and the quantity of water and the river slope". He recognized four "phases" from source to sea:

1 The mountain or torrent phase. This is characterised by steep slopes, falls and gorges, and receives an abundant supply of course, angular sediment. Such streams and rivers are widespread in the mountains of New Zealand (Figure 16.2A).

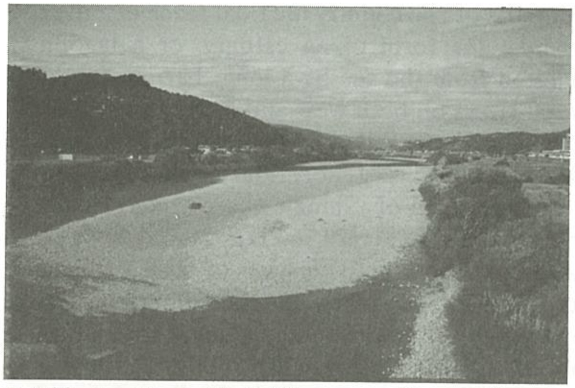
2 The shingle phase. Where the river leaves the mountains, it flows in shallow braided channels which wind across a wide gravel bed. The course tends to be straight, with a steep slope; the river may be aggrading, and probably reworking earlier deposits as it erodes its banks and migrates laterally. The rivers flowing across the Canterbury Plains are good examples, as is the lower Hutt River (Figure 16.2B).

3 The silt phase. In some rivers, there is an abrupt transition to a channel with sand bed and silt banks. The relatively narrow, deep, and low gradient channel of the Waipaoa River near Gisborne is a good example (Figure 16.2C).

4 The tidal phase. This class of river is rather restricted in distribution in New Zealand, with only the lower few kilometres of most rivers being included. The tidal phase of most rivers is very similar to the silt phase, with a deep, narrow, sinuous channel composed of silt banks and sand bed. The best examples are found in Northland; in

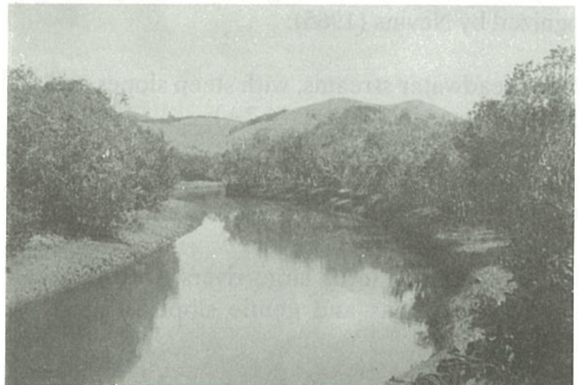


A Mountain or torrent phase: Roaring Burn, Fiordland.



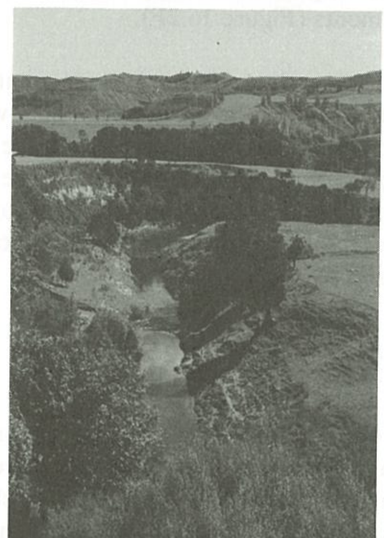
B Shingle phase: Hutt River near Lower Hutt.

C Silt phase: Waipaoa River near Gisborne.



D Tidal phase: Whangaroa Harbour, Northland.

E Braided river: Waimakiriri River near Arthurs Pass under low flow conditions during winter.



F Entrenched channel: Turakina River near Wanganui.

Figure 16.2 Representative Channel Types.

other parts of the country the tidal reach of a river may take the form of an estuary, or a lagoon separated from the sea by a long barrier beach. Several rivers on both west and east coasts of the South Island - the Rakaia, Waimakariri, Hokitika, or Karamea have very short tidal reaches immediately downstream of a shingle phase which extends almost to the sea (Figure 16.2D).

A more recent classification of New Zealand rivers used cluster analysis of data describing the morphology, hydrology, and sediment load of almost 200 rivers (Mosley, 1987). Four distinctive classes of river were revealed, the first three of which match mountain, shingle, and silt phases recognized by Nevins (1965):

1 small headwater streams, with steep slopes and coarse bed material (Figure 16.2A);

2 braided rivers, with gravel beds and large values of width relative to depth (Figure 16.2E);

3 the lowland course of large rivers, with large cross-sectional areas and gentle slopes (Figure 16.2C);

4 entrenched channels with narrow widths relative to depth, commonly located in mudstone catchments (Figure 16.2F).

However, these four groups accounted for only about 20% of the rivers; the other 80% were broadly similar to each other, and tend to be of the type designated as shingle phase (Figure 16.2B) by Nevins (1965). Hence, it is difficult to develop a useful morphological classification of New Zealand rivers.

Braided and Meandering Rivers

The obvious differences between braided and meandering rivers have been extensively studied. A single river may have a variety of channel patterns along its course, and may change from braided to meandering or vice versa.

Several studies have attempted to discriminate between braided and meandering rivers on the basis of discharge, channel slope, and median grain size (Leopold et al., 1964; Carson and Griffiths, 1987a, pp 128-134). Analysis of the morphology of rivers in the Canterbury Plains (Carson, 1984a, 1984b) indicated, however, that braided and meandering rivers cannot be separated on this basis, and that other factors, such as bedload supply, transport rate, stream power, and bank erodibility control them. Carson (1984b) classified channel patterns of New Zealand gravel-bed rivers, using bank erodibility and relative bed material supply rate. Nevertheless, the implication of Mollard's (1973) classification (Figure 16.1), that rivers actually form a continuum, should be borne in mind.

Processes of Channel Formation

The Stream Bed

As discharge in a river increases, so too do depth, velocity, and shear stress "$\tau = \gamma.R.S$", in which gamma is the unit weight of water, R is hydraulic radius, and S is the water surface slope. At some point, velocity exceeds the threshold at which sediment on the stream bed can remain stable, and sediment transport is initiated (Chapter 13). Erosion, transport, and deposition of bed and bank sediments are the principal processes whereby stream channels are modified by flowing water, as sediments are set in motion in one place and redeposited in another, and bedforms like bars, riffles and pools migrate and reform.

Bedforms in channels form and migrate in a complex manner, particularly in braided and transitional gravel-bed rivers. Bars may be constructional (formed by deposition of sediment) or erosional (as newly forming channels dissect old bars, creating new residual features).

The most common is the riffle (Figure 16.2B), also known as a diagonal bar, skew shoal, and various other terms. The riffle generally looks like a steep-faced bank of gravel, stretching diagonally across the channel from one bank to the other. At low flow, water flows rapidly over the riffle because its gradient is greater than the pools up and downstream, but at high flow the riffle becomes "drowned out" (Figure 16.7).

A basic building block in braided gravel rivers appears to be the "confluence-chute-bar" (Davoren and Mosley, 1986) or "fishscale bar" (Carson and Griffiths, 1987a). In this system, lateral erosion of bars upstream and scour in the chute itself provide sediment which is transported through and redeposited on the bar downstream, where flow is diverging and losing its competence to carry sediment (Figure 16.3A). Eventually, flow tends to concentrate to one side or the other of the

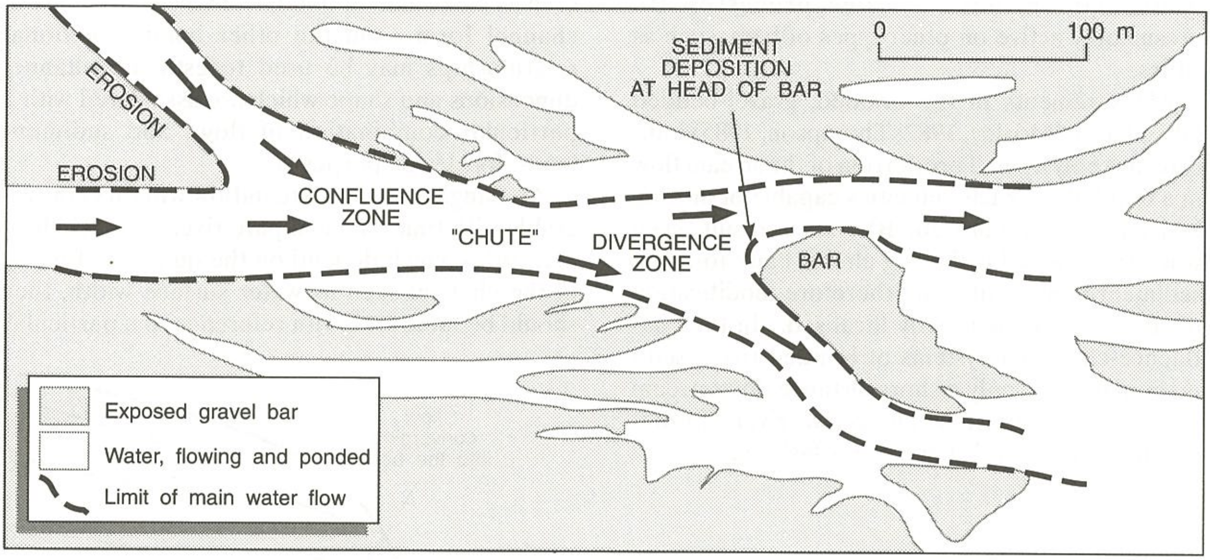


Figure 16.3(A) A confluence-chute-bar system in the Ohau River (Davoren and Mosley, 1986).

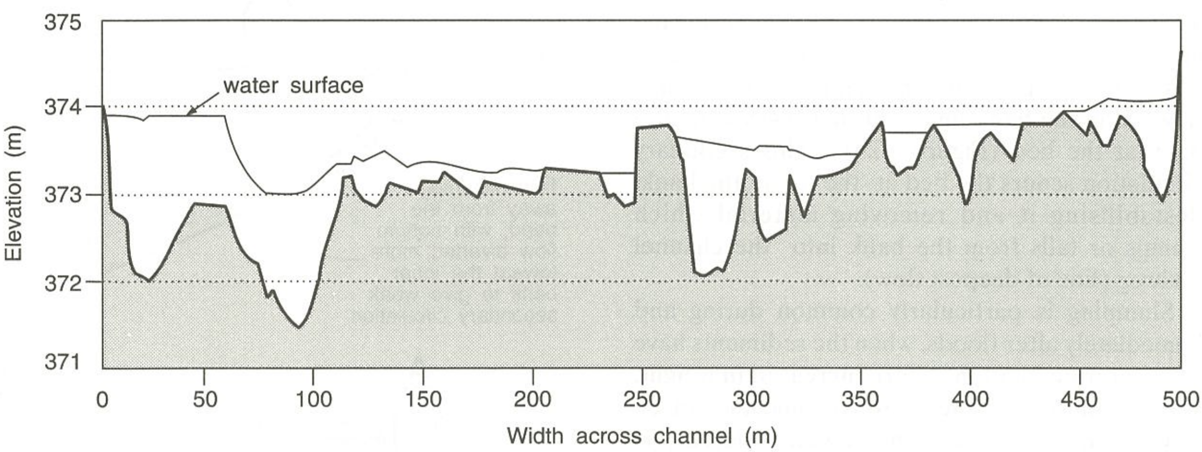


Figure 16.3(B) A section surveyed across the Ohau River when it was flowing at 500 m3/s, approximately bankfull discharge. The water surface profile is very uneven across the channel, and water depths and velocities are great enough only in the four deepest channels (at 20, 90, 280, 490 m) for sediment to be transported.

downstream bar, cutting laterally into the sediments, and recommencing the process in a different place. Hence, sediment moves in relatively short steps, and may be stored for long periods, if it is deposited in a bar which becomes inactive as the channels migrate away from it. The symmetrical confluence-chute-bar in Figure 16.3 is relatively uncommon, but the same processes are presumably active on other types of bars, such as riffles.

Measurements in the braided, gravel-bedded Ohau River (Mosley, 1982; Thompson, 1985) indicate that only a small proportion of the stream flow in a braided river has velocities capable of moving bed material (Figure 16.3B). As a result, even when the channel is flowing almost bank-to-bank, sediment movement - and therefore modification of the channel - is usually localised. In the confluence-chute-bar systems of braided rivers sediment may move only a short distance downstream before it is redeposited. Hence, the river's form is modified in a very discontinuous fashion.

The Stream Banks

Channel shape is determined in part by bank erosion. It is commonly greatest on the outer bank in a bend, against which flow is concentrated, and where flow depths are greatest. The water flowing into a bend tends to pile up against the outer bank, setting up intense turbulence and a net downward flow along the bank line, and flow away from the bank at the bed (Figure 16.4). This secondary circulation scours the bed at the toe of the bank, destabilising it and removing material which slumps or falls from the bank into the channel thalweg (line of deepest flow).

Slumping is particularly common during and immediately after floods, when the sediments have been saturated and therefore increased in weight. Where cohesive sediments overlie sand and gravel, which is more readily washed away, undercutting may lead to collapse of large blocks into the channel. Banks may also be eroded by the direct force of water during high flows, by drop impact, rain and by frost action during low flow when banks are exposed to the weather.

Channel Morphology

The functional relationships between streamflow, stream channel form, drainage network geometry, and watershed characteristics may be expressed in terms of mathematical or statistical models. These can be used to estimate hydrological parameters such as mean annual flood, from measurements of channel form. On the other hand, functional relationships may be used to estimate channel dimensions and shape which are associated with a particular combination of flows and sediment loads, for design purposes.

At a single location, streamflow will vary considerably with time. To compare river channel characteristics which depend on the quantity of water in the channel such as water surface width, they should be measured with reference to a particular

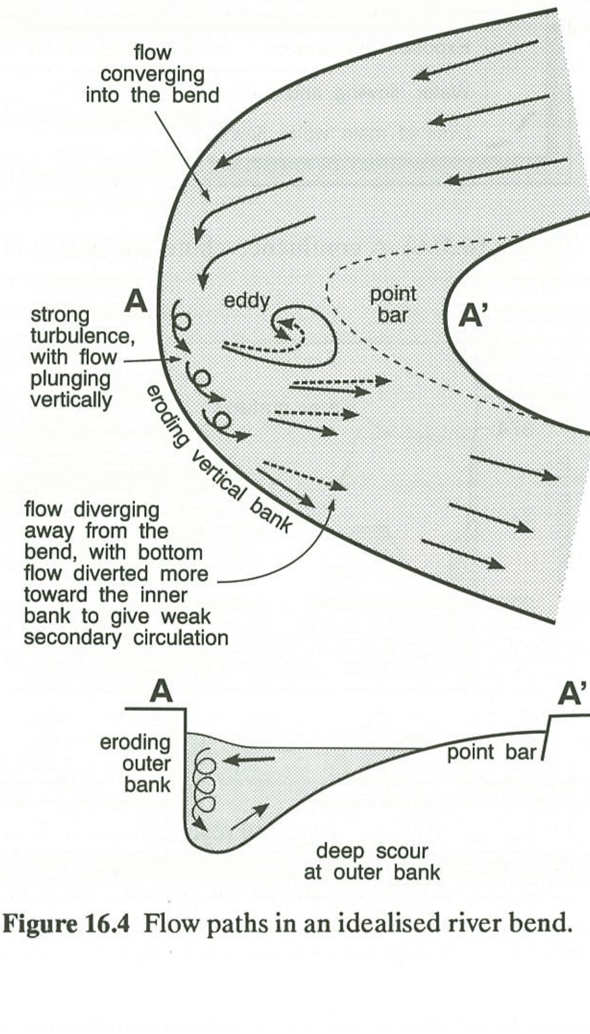


Figure 16.4 Flow paths in an idealised river bend.

index of streamflow, for example mean discharge (Q bar) or mean annual flood ($Q_{2.33}$). The latter is the average of the series of annual maximum flows, which in New Zealand has an average recurrence interval of 2.33 years.

Many studies have used as a reference the bankfull discharge, the discharge which just fills an alluvial channel and starts to inundate the floodplain. The bankfull discharge is often taken to approximate the mean annual flood. In a sample of 72 South Island rivers, however, bankfull discharge had a recurrence interval between less than 1 and more than 10 years, with a median of about 1.5 years (Mosley, 1981a).

Bankfull discharge is frequently considered to be equivalent to the dominant discharge, that is, the discharge which is most effective in transporting sediment and reshaping the channel bed and banks (Figure 16.5). The area under curve C in Figure 16.5 can be readily integrated, using a computer, and the effects of varying shapes of curves A and B therefore taken into account more satisfactorily than by using a single index value.

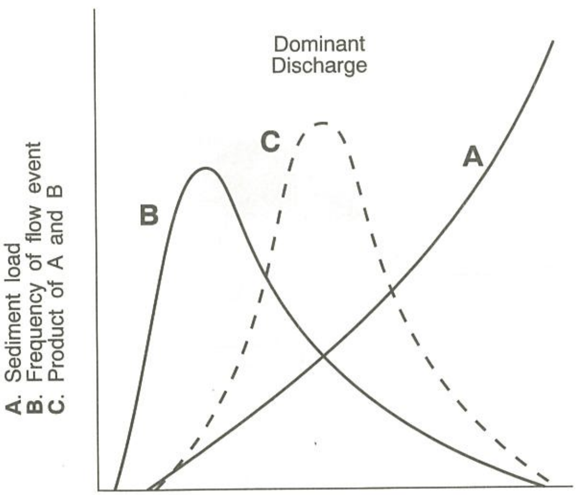


Figure 16.5 Definition of dominant discharge in terms of sediment transported by a range of discharges (after Wolman and Miller, 1960).

The concept of bankfull discharge as the dominant or channel-forming flow is not appropriate for all channels, particularly those with bedrock banks, braided channels, or those which are known to be actively aggrading or degrading. Such channels are, of course, very common in New Zealand.

River channels can be described in terms of their cross-sections at one location along the channel; their reaches (lengths of channel which can for a particular purpose be treated as similar throughout); and their channel network in an entire catchment. Many morphometric indices have been developed; selections are provided in Goudie (1981), Gregory and Walling (1973), and Richards (1982). Many are duplicative, have been rarely used, or have little application for hydrological purposes. The more important are summarised in Table 16.1.

The Channel Cross-section

The measurements used to describe channel cross-sections are based on standard land surveying techniques or streamflow measurement (Figure 16.6).

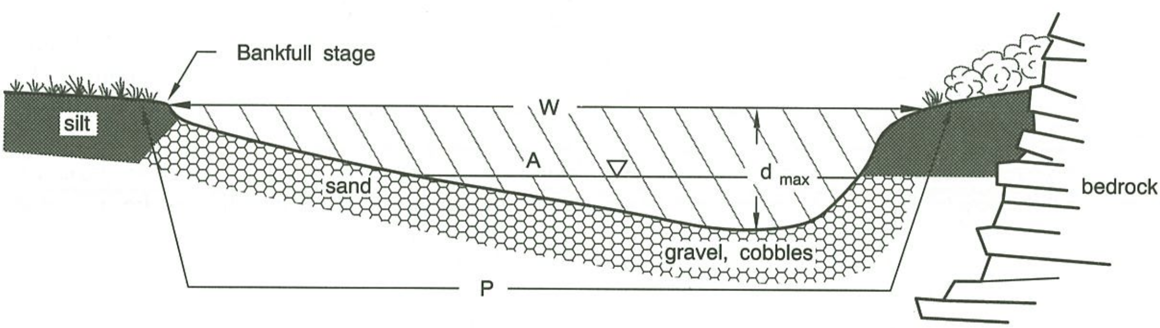
Cross-section date are variable because of the differences between the dimensions of features such as riffles and pools (Figure 16.7). Width and depth also vary considerably along channels (Beschta, 1983; Knighton, 1981; Mosley, 1981b). Some variations repeat more or less regularly in association with riffles, bends, and pools; others are virtually random, reflecting the occurrence of bedrock outcrops, fallen logs, and so forth (Figure 16.8).

For comparison, cross-sections must be selected in a consistent fashion, for example by always choosing a cross-section at the cross-over between two bends, or by making measurements at a number of sections in each reach, and calculating the mean and standard deviation of each morphometric variable. Up to 20 to 30 sections may be required to accurately characterise a given reach.

Channel dimensions such as width tend to remain constant between tributaries, other things being equal. On the larger scale, they increase from source to sea as tributaries add more water to the river (Figure 16.12). The reach of Cave

	Channel Cross-section	Channel Reach	Network	Drainage Basin
Length or scale measures	Elevation Width, wetted perimeter Mean depth, hydraulic radius Maximum depth Evaluation	Meander wavelength Bend amplitude Radius of curvature Bedform wavelength Bedform amplitude Width, wetted perimeter Mean depth, hydraulic radius Maximum depth	Total channel length Drainage density (length per unit area) Network diameter Mainstream length	Perimeter Length Outlet elevation Maximum elevation Relief
Area or Extent	Channel capacity (cross-sectional area)	Channel capacity (cross-sectional area) Area of riverbed	Stream order, magnitude Number of nth order streams	Area Area in lakes, ponds, wetlands Area under forest cover,
Shape	Form ratio (width/mean depth) Depth ratio (maximum depth/mean depth) Roughness (Manning n, Chezy C, Darcy-Weisbach, f) Sediment size (d_{50})	Sinuosity Braiding index	Bifurcation ratio	Elongation Circularity
Angular or gradient	Energy gradient Water surface slope Transverse water surface slope	Energy gradient Water surface slope		Relief ratio

Table 16.1 Morphometric Indices Of Waterways And Drainage Basins



KEY

W Total channel width at bankfull
A Cross-sectional area at bankfull
P Wetted perimeter at bankfull
R Hydraulic radius = A/P
d Mean depth = A/W
d_{max} Maximum bankfull depth

Figure 16.6 Definition sketch of a channel cross-section.

Stream shown in Figure 16.8, which flows from the Craigieburn Range in Canterbury, is unusual in that its width decreases in the downstream direction (Knighton, 1981). This may be in response to a downstream readjustment to the abrupt effect of a large tributary at the top end of the reach.

The Channel Reach

Gradient determines water velocity, which in turn governs the time that water takes to travel down a river, and the stream's ability to transport sediment. In streams and small rivers gradient can vary over short distances, so field survey is needed. For larger rivers, flood-plain gradients may be measured from 1:50,000 maps; over distances greater than 1 or 2 wavelengths of riffles or bends, it approximates water surface slope or energy gradient, if referenced to equivalent points on the repeating riffle-pool sequence.

Channel pattern can readily be measured from topographic maps and aerial photographs. There is a strong statistical relationship between meander wavelength and discharge, although few rivers have regular repeating meanders. Meander wavelength commonly increases as the 0.5 power of mean annual flood. Sinuosity, the ratio of distance along the channel to straight line distance, is related to sediment and topographic characteristics and to the energy of a waterway, rather than to indices of discharge (Figure 16.1). The relationships between the planimetric charac-

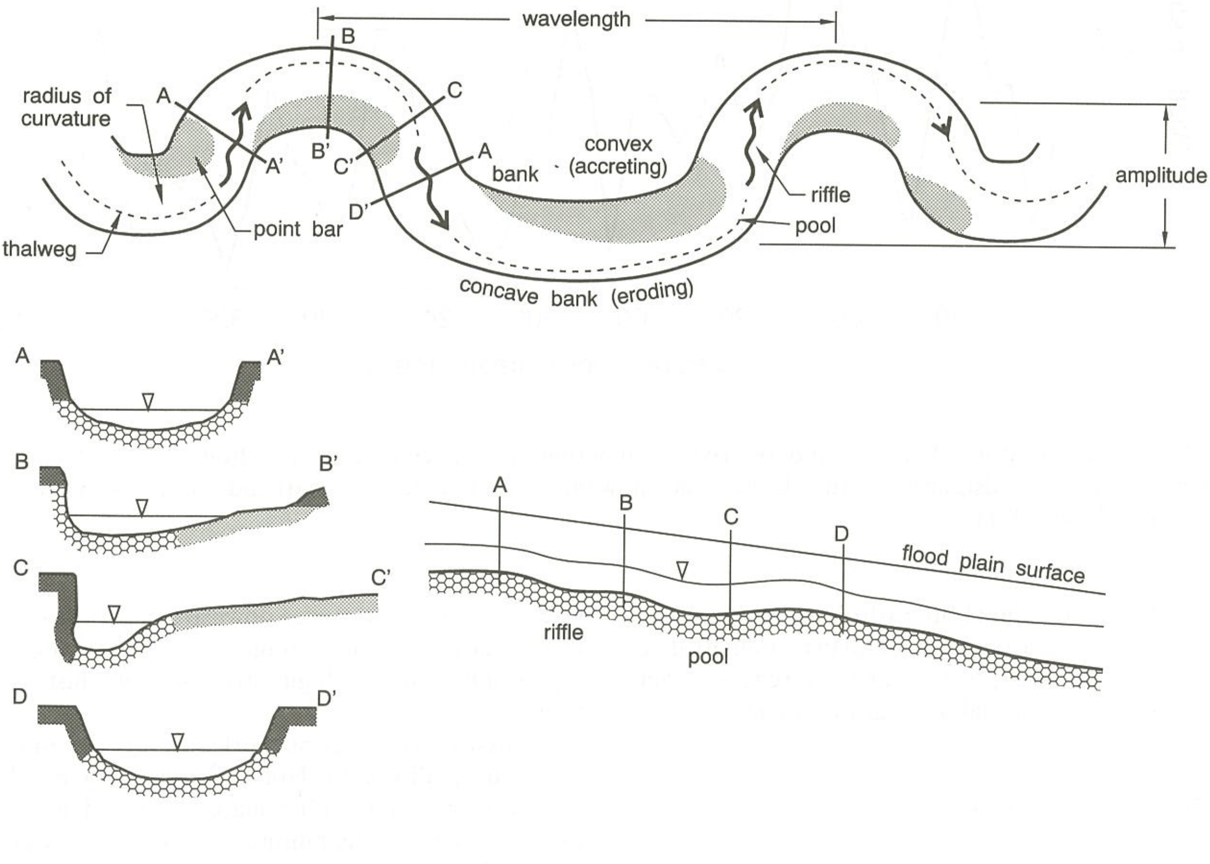


Figure 16.7. Idealised plan geometry of a meandering river.

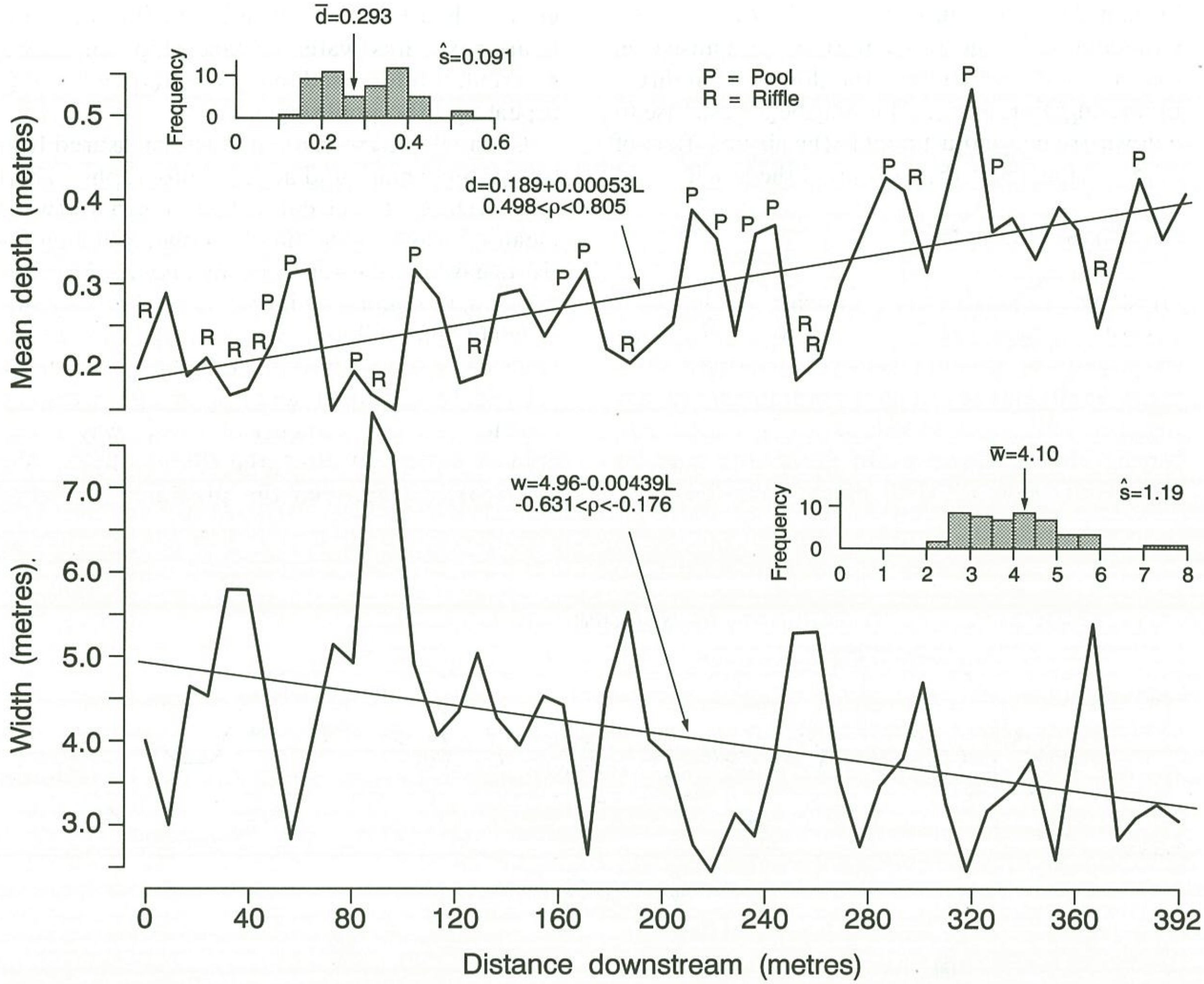


Figure 16.8 Width (w) and mean depth (d) as a function of distance L along a section of Cave Stream, Canterbury. The histograms of the observations show the average values ($\bar{w}$ and $\bar{d}$) and standard deviation ($\hat{s}$) (Knighton, 1981).

teristics of a channel, its hydrology, and its sediment load are particularly useful in reconnaissance studies, for example to make inferences about hydrology from aerial photographs of a river.

The Channel Network

Channel networks both reflect hydrologic processes and partially control them, by controlling the rate at which water moves downstream. Network characteristics also strongly reflect lithology, topography, and geologic structure and history (Gage, 1980).

Analyses of drainage networks are very dependent on the quality of the data source, and in general only modern topographic maps prepared using photogrammetric techniques, at scales greater than 1:50,000, can be confidently used. The methods used in defining the extent of the channel

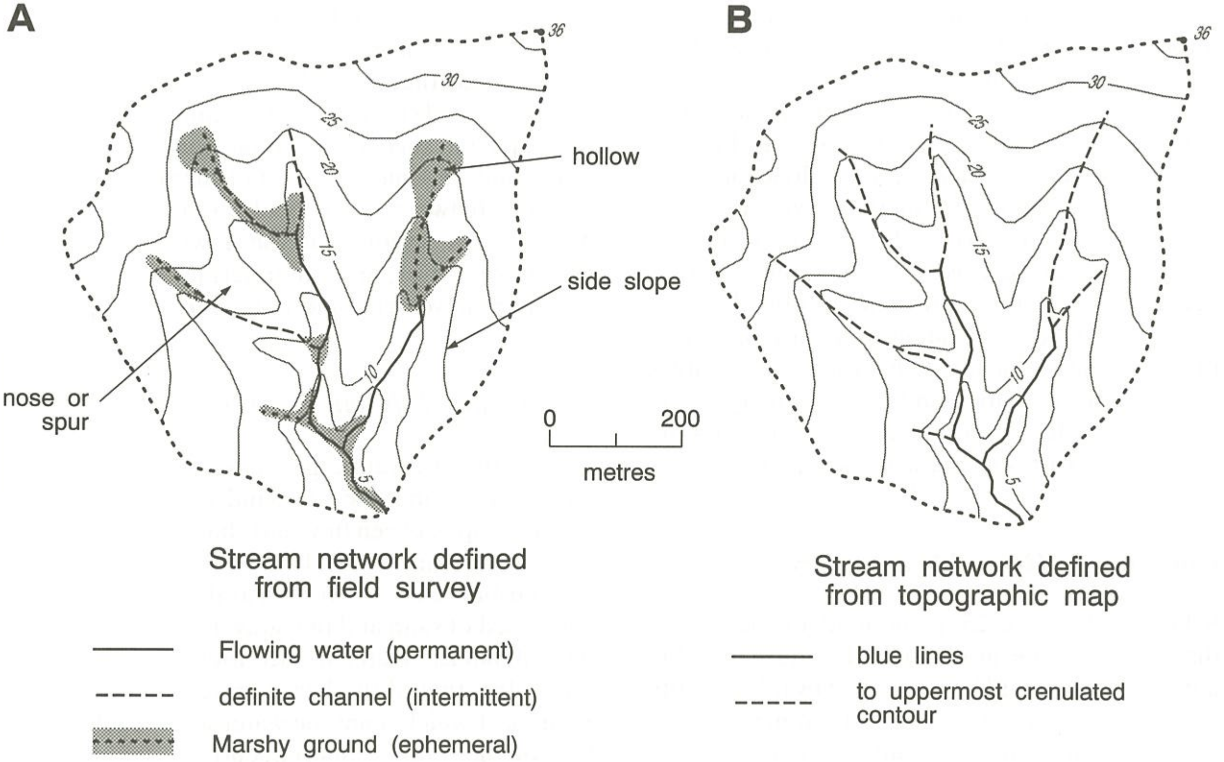


Figure 16.9 Definition sketches of channel networks, for an imagined catchment. Network defined from field survey (A), and from a topographic map (B).

network, particularly the identification of fingertip tributaries and the location of their sources, must be consistent if intercomparisons are to be made. Marked streams on topographic maps normally delimit "normal summer flows"; networks can be extended into topographic hollows, marked by contour crenulations, on the assumption that surface flow during rainfall events will extend into locations where saturation occurs (Figure 16.9).

When a network is being surveyed or checked in the field, several conventions may be adopted, for example: (1) channels with flow during the survey; (2) defined permanent channels, whether or not flowing at the time (ephemeral streams); and (3) topographic hollows with surface saturation or with vegetation characteristic of marshy ground. In some areas, the stream may extend even further, in the form of subsurface pipes or tunnel gullies. Streamflow generation by rapid subsurface flow through macropores in the soil makes even more difficult the distinction between streamflow and its surface expression as stream channels, and subsurface flow processes.

Once a drainage network has been mapped and defined, a variety of geometric and topologic attributes may be defined (Table 16.1). The topologic characteristics may be described by the stream ordering systems developed by Strahler (1952) and Shreve (1966). Horton's (1945) "laws" of drainage network composition relate stream numbers, catchment areas, and lengths to stream order. In the Strahler system of stream ordering, the "fingertip" tributaries are assigned order 1; two 1st order channels combine to become, below their confluence, a 2nd order channel, and so on. The junction of a lower order channel (eg a 2nd order flowing into a 3rd order) does not change the order of the higher order stream. In the Shreve system, the fingertip tributaries are assigned magnitude 1; each downstream channel segment has a magnitude equal to the number of 1st magnitude tributaries upstream.

Drainage basin area is perhaps the easiest characteristic to relate to hydrology. Indices of streamflow such as mean annual flood or mean annual runoff generally increase with drainage area (Figure 16.10), although storm-period specific discharge (discharge per unit area) is generally inversely proportional to area, because of the restricted areal extent of intense precipitation. Useful estimates of hydrological variables generally must combine indices of drainage basin or network shape with indices of precipitation, vegetation cover, soil characteristics, etc.

Controls on River Morphology

Schumm (1969, p. 256) pointed out that "the dimensions, shape, gradient, and pattern of stable alluvial rivers should be controlled by the quantity of water and quantity and type of sediment moved through their channels", and summarized these relationships in a series of conceptual equations. Hydraulic geometry relations for rivers throughout the world demonstrate a high degree of consistency in the way that they adapt to stream flow, sediment load, and conditions which limit free channel development, such as inerodible banks. The basis of hydraulic geometry is a set of simple power law relations between an index of water discharge, as it varies at-a-station or in the downstream direction along a river, and water surface width, mean depth, mean velocity, and water surface slope.

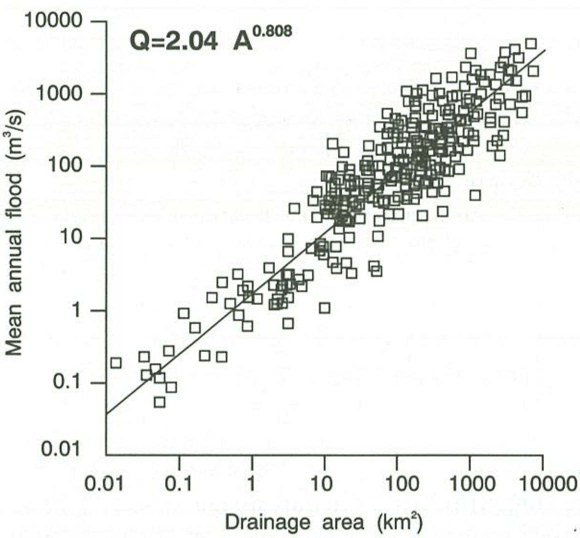


Figure 16.10 The relationship between mean annual flood and drainage area for New Zealand catchments.

At-a-Station Hydraulic Geometry

At-a-station hydraulic geometry plots for some typical rivers in New Zealand demonstrate the relationship between flow and channel dimensions and shape (Figure 16.11). The Waipaoa River near Gisborne has a broadly parabolic shaped bed composed of sand and fine gravel, and almost vertical silt banks. Water surface width increases as a power function of discharge up to flows of about 30 m^3/s, at which point the water level reaches the foot of the banks. Subsequently, water surface width increases more slowly - but still as a power function of discharge - as discharge increases. The plot for mean depth as a function of discharge similarly has two segments which reflect the shape of the channel, but mean velocity increases as a simple power function of discharge across the full range of flows.

The Buller River near Murchison, which has a rock-bound channel with an almost flat alluvial bed, shows a set of relationships similar to that for the Waipaoa. The Buller has rock banks, and is significantly narrower at the same discharge than the Waipaoa, with silt banks. For example, at flows of 100 m^3/s the Buller is 47 m wide, while the Waipaoa River is 87 m wide. Their depths are similar, about 1.9 m but the mean velocity of the Buller is higher - 1.6 m/s as against 1.1 m/s, and increases more rapidly as discharge increases.

Before its diversion for the upper Waitaki power scheme, the Ohau River was a broad, gravel-bed braided river with non-cohesive gravel banks and multiple channels. The average width of flowing water, summed across all channels at several

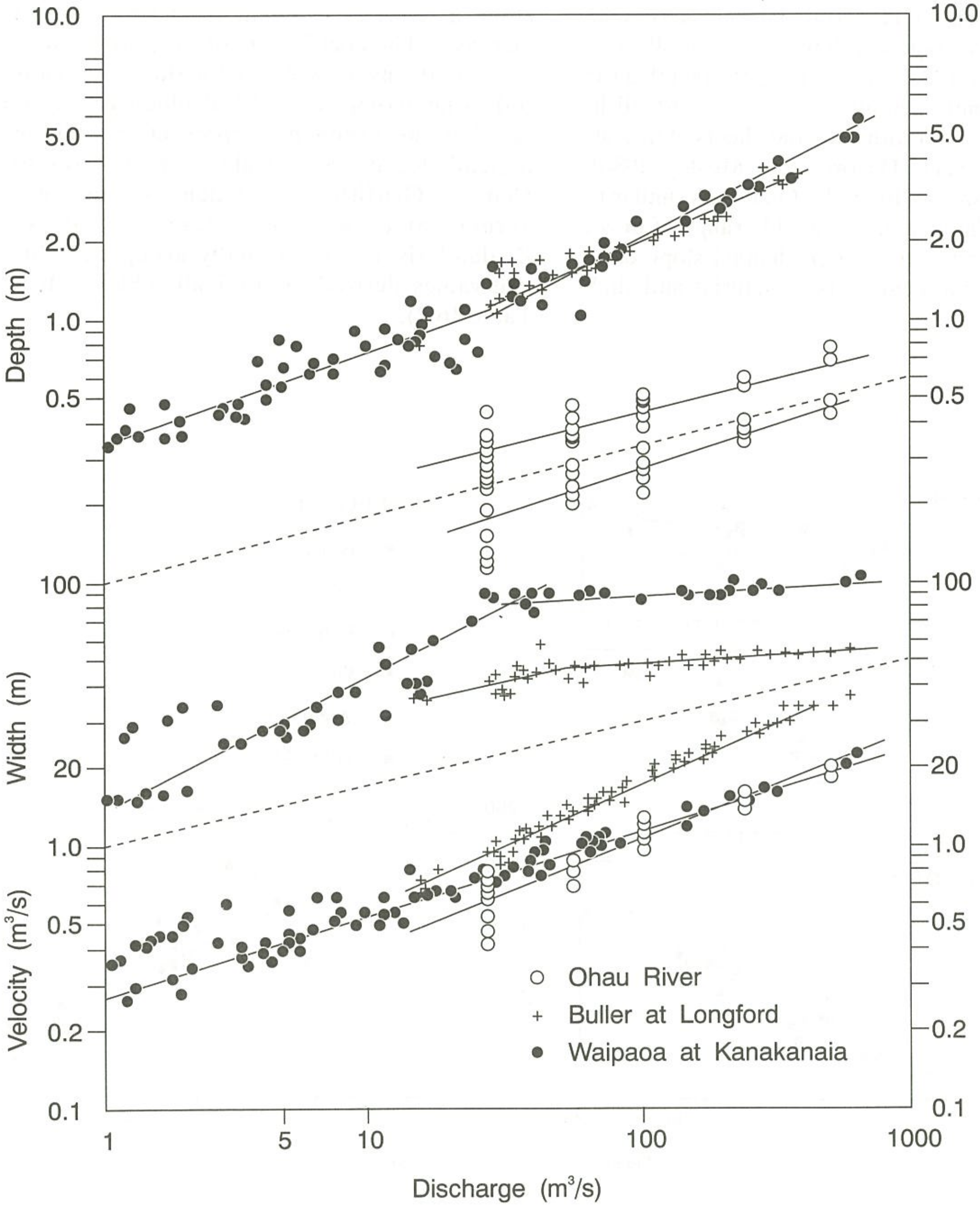


Figure 16.11 At-a-station hydraulic geometry plots for the Buller, Ohau, and Waipaoa Rivers.

cross-sections, was much greater in the Ohau than in the Waipaoa or Buller, about 250 m at a discharge of 100 m^3/s, reflecting the lack of cohesion of the Ohau's banks, and the ease with which branch channels could migrate laterally across the gravel bed. Average depth was substantially less, about 0.45 m at 100 m^3/s, reflecting the fact that the branch channels commonly had a broad, dish-shaped form, often with wide flat sheets of migrating gravel present (Davoren and Mosley, 1986). However, flow velocity in the Ohau was similar to that in the Waipaoa across a wide range of flows, probably because its greater channel slope compensated for the coarser bed material and shallower depth.

Downstream Hydraulic Geometry

Similar relationships between discharge and channel dimensions are observed downstream along a river, as drainage area and discharge increase. The coefficients of proportionality of the equations vary from locality to locality, reflecting geological and hydrological differences, but the exponents represent the physical dependence of the variables upon water discharge. Griffiths (1980) demonstrated good agreement between exponents for six New Zealand rivers, traditionally accepted values, and values derived theoretically (Figure 16.12; Table 16.2).

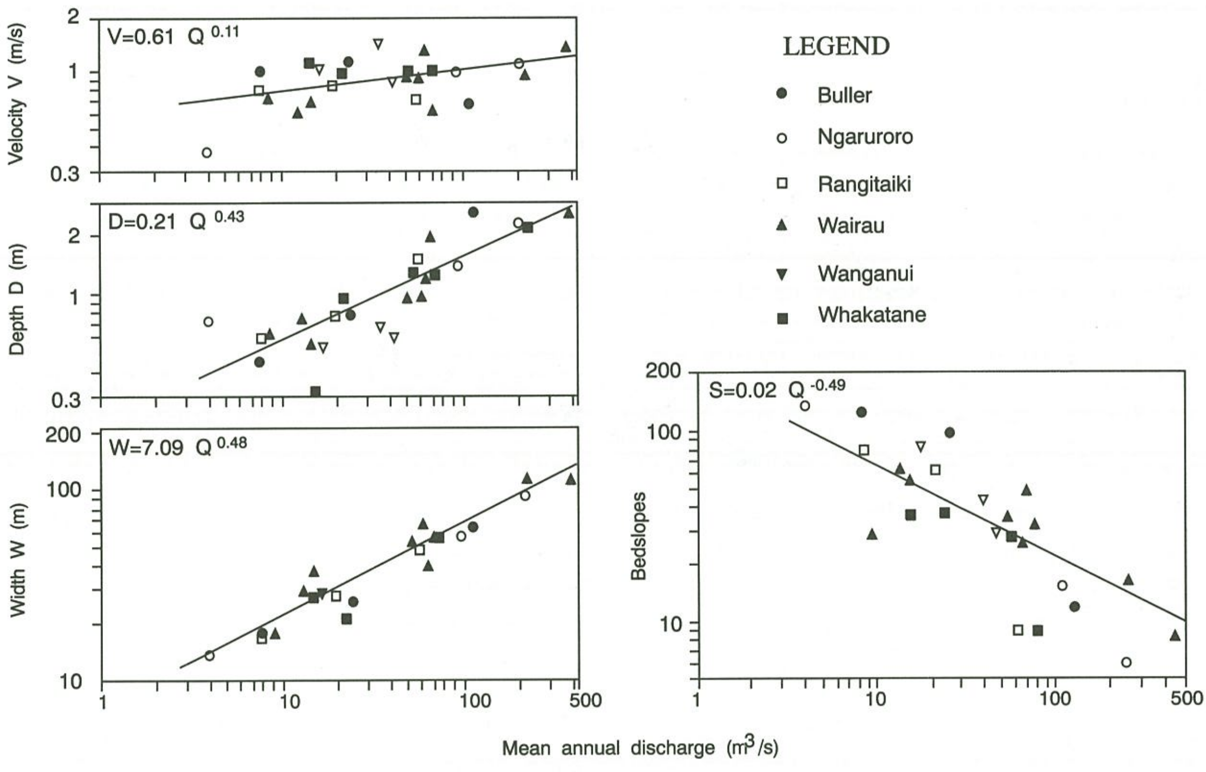


Figure 16.12 Downstream hydraulic geometry relationships along six gravel-bed rivers (Griffiths, 1980).

Variable	Relation	Exponent		
		Traditionally accepted value (bankfull or mean annual water discharge)	Field value for NZ gravel bed rivers (mean annual water discharge)	Derived theoretical value (bankfull water discharge)
Water surface width	$W \propto Q^b$	0.5	0.48	0.44
Mean flow depth	$d \propto Q^b$	0.4	0.43	0.44
Mean flow velocity	$V \propto Q^b$	0.1	0.11	0.11
Suspended sediment concentration	$C \propto Q^b$	-0.2*	0.31	-0.33
Suspended sediment load	$G \propto Q^b$	0.8	1.31	0.67*
Energy gradient	$S \propto Q^b$	−0.5	−0.49	−0.44
Darcy-Weisbach friction factor	$f \propto Q^b$	−0.3*	−0.38	−0.22
Manning coefficient	$n \propto Q^b$	−0.3	−0.07*	−0.04*

* denotes exponent derived from definitions, $C_m \propto C_m Q_m$: $f \propto d_m S_m / V^2_m$: $n \propto d_m 0.66 S_m 0.5 / V_m$; using values listed.

Table 16.2 Values of Exponents in the Relations for Downstream Hydraulic Geometry of River Channels (Griffiths, 1980)

The Effect of Variables other than Discharge

Factors other than discharge also control channel form. Griffiths (1981, 1983) developed a regime equation for the width of a stable alluvial channel at near-threshold conditions for sediment transport:

$$w = 5.28\, Q\, s^{1.26} / d_{50} 1.5, \qquad (16.1)$$

in which d_{50} is median diameter of the surficial bed material. His equation 16.1, based on analysis of flow resistance and conditions for sediment entrainment, indicates that channel width is dependent also on valley slope and sediment size.

A study of the hydraulic geometry of 72 rivers in the South Island indicated that only 53% of the morphological differences between rivers could be statistically "explained" by variables describing river flow, sediment character, bank erodibility, and flow variability (Mosley, 1981a). For some individual attributes of river form, the level of explanation was much higher; for instance, nearly 90% of the differences between rivers in cross-sectional area at bankfull stage could be accounted for by differences in mean annual flood, mean diameter of bed sediment, and the percentage of silt and clay in the banks (an indicator of bank erodibility).

The 47% of the difference between rivers which was not "explained" by these variables must be accounted for by other factors which affect the resistance of the bed and banks to erosion, such as bedrock outcrops and vegetation along the banks. River engineering practice in New Zealand has for decades relied on replicating these natural constraints on bank erosion, by employing riprap, groynes, and willow plantings (Acheson, 1968).

The Effect of Extreme Events

Wolman and Miller (1960) pointed out that the low flows which are experienced for much of the time do not have the energy required to transport sediment or erode banks, whereas extreme or catastrophic events - even though they may cause dramatic changes in the landscape - happen too infrequently and for too short a period to accomplish a significant proportion of the total work expended in shaping the landscape. Instead, it is the moderate events with a return period of months to a year or two which accomplish most geomorphic work, in total. In New Zealand, where the evidence of extreme events is common, this proposition my have less validity than in parts of the world which are geologically and climatically less active (eg Beschta, 1983; O'Loughlin, 1969; Pain, 1968; Phillips, 1988).

There are many locations in New Zealand where extreme events have had a dramatic effect on the rivers, and Pain (1968) argued, from observations in the Hunua Ranges, that extreme events may have particular significance in New Zealand, when mass movement and gullying on hillsides supply large quantities of sediment to channels. An extreme case is where landsliding creates landslide-dammed lakes, of which there are many in New Zealand (Adams, 1981).

The particular sequence of flow events down a channel controls variations about its average state. Large, infrequent floods may cause dramatic changes in a particular section or reach of channel, which may not return to equilibrium with more moderate "channel-forming" flows for some time.

For example, a storm in April 1951 triggered mass movements and numerous debris flows in the Kowai River and its tributaries, Canterbury, (Beschta, 1983). The active channel in many places aggraded and widened to fill the entire valley bottom. Subsequent reworking of the deposits has caused degradation and narrowing, with degradation by up to 6 m in the Foggy River between 1951 and 1982, and scouring of over 500 m^3 of gravel per metre of channel length (Figure 16.13).

Beschta (1983) estimated that a minimum of 295,000 m^3 of sediment was transported from the Foggy River into the Kowai River during 1951-1982. There has been a downstream progression of depositional zones, as gravel has moved - almost as a "wave" - and been redeposited during subsequent storms. Hence, the 1951 storm initiated a sequence of channel adjustments which has continued for several decades, during which channel shape and dimensions were not in equilibrium with more moderate flows.

The Influence of Human Activity

The factors which control river channel shape and behaviour such as flows, sediment load and size, bank vegetation and erodibility, can be modified by human activity, leading to changes in channel shape. Such changes may be unintended, as in the

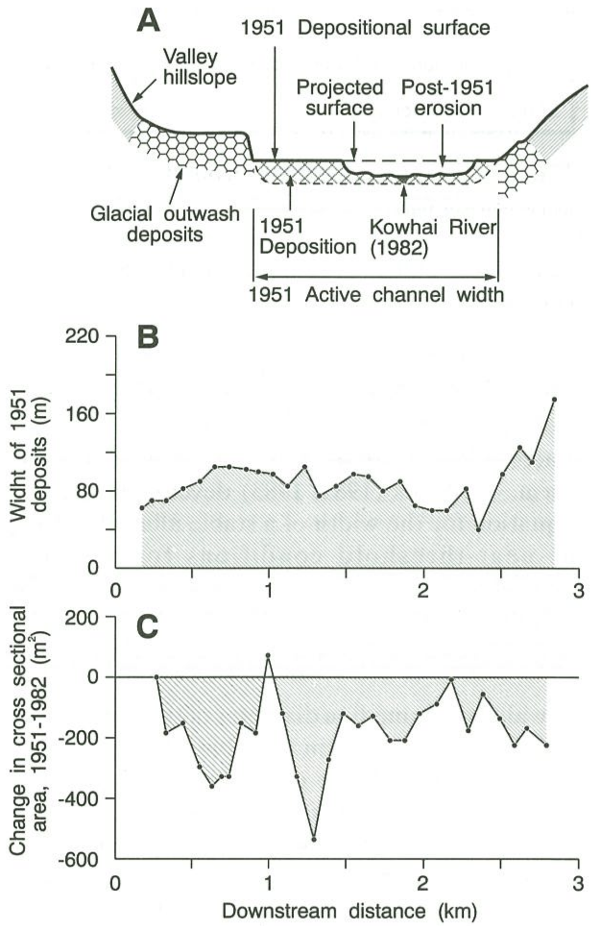


Figure 16.13 Channel changes in the Foggy River, Canterbury, in response to a major storm in 1951. Schematic cross-section (A); width of fresh 1951 deposition (B); erosion of deposits during the period 1951-1982 (C). (Beschta, 1983).

case where forest clearance leads to increases in sediment load and consequent aggradation and channel widening, or intentional, as where rock riprap or planted willows are designed to prevent bank erosion and reduce channel width.

Flow Modification

Flows in many New Zealand rivers have been modified, principally by abstraction for irrigation, water supply, and power generation. For example, the mean flow of the Waiau River has been reduced by the 270 m^3/s which has been diverted to Doubtful Sound via the Manapouri Power Station.

Less commonly, there has been flow augmentation, notably in the case of the Waikato, which receives water diverted through the Tokaanu power station from the Whanganui and Whangaehu systems.

Rates of abstraction may be a significant proportion of the commonly occurring low to medium flows, even to the extent of entirely drying up a channel, except perhaps for intermittent pools. This clearly has a major impact on the morphometry of the water body in the channel, and on its value for fish habitat, recreation, and other instream uses (Mosley, 1983). However, it may not affect the shape of the channel itself, if - as is commonly the case - abstraction is a small proportion of the flood flows which are principally responsible for moving sediment and shaping the channel.

Such effects have been extensively studied overseas (Petts, 1984), but to a limited extent in New Zealand. The rivers particularly affected by major flow modification tend to have hydrologic regimes which are already modified by lake storage, notably the Waikato, Waiau, and Waitaki, so that effects on channel morphology could be expected to be limited or unimportant. On the other hand, some rivers from which all flows have been diverted - the Ohau and Pukaki - no longer exist except as spillways, and hardly need study to demonstrate the impacts on channel morphology.

A more indirect effect of human activity on river flows is through the medium of land use and vegetation change. Chapter 15 discusses how conversion of forest to pasture, or afforestation of pasture, may affect streamflows. Generally, low flows are modified to the greatest extent, and floods to a decreasing extent. The significance of such changes in streamflow for channel morphology are, therefore, likely to be similar to those resulting from flow abstraction.

Sediment load

Sediment loads in many New Zealand rivers have been affected by human activity. Direct effects result from gravel and sand extraction, which has led to locally severe degradation and bank instability in rivers such as the Manawatu and Otaki, and to disposal of spoil directly into channels, for instance from the construction of logging roads and landings, or clearance of landslide debris from the public roading system.

A more widespread effect results from changes in land erosion rates and sediment supply to rivers, as a consequence of vegetation modification, land use change, and land management practices (chapters 13 and 15). There are many case studies of such impacts, including those associated with removal of forest from stream banks, and introduction of large volumes of sediment into rivers as a consequence of major landsliding following deforestation.

For example, the Tamaki River, which drains from the eastern Ruahine Range, responded rapidly to forest clearance along its banks in the early years of this century. In the first 2.5 km from its exit from the Range, it widened from less than 10 m in the early 1920s, only a few years after forest clearance, to an average 54 m in 1942 (Mosley, 1978). From 1942 to 1976, channel width increased by only another 6 m. The widening during the earlier period, which was accomplished particularly during large storms, was probably a response to declining bank resistance as tree roots decayed.

Perhaps the greatest impacts of vegetation and land use change have been in the East Cape region, notably the Waipaoa River catchment. It has been estimated (Poole, 1983, p 69) that, as a result of accelerated erosion in the headwaters, 30 to 40 million cubic metres of silt and gravel have been stored in the middle reaches of the river, causing substantial aggradation and channel widening.

River Control

Deliberate modification of river channels has been undertaken on an extensive scale in New Zealand for many decades, to mitigate the effects of bank erosion, bed aggradation, and flooding. Often, these efforts have in effect reversed the consequences of other human activity. For instance, removal of forest from floodplains and river banks to enable pastoral agriculture has commonly had the unwanted side effect of accelerating bank erosion rates. The response has been to plant willows along thousands of kilometres of river bank to replace the trees whose roots formerly resisted bank erosion.

Where planted willow poles do not give sufficiently effective or rapid protection, other means such as rock riprap and the construction of groynes are used, for example along the Waimakiriri River. Riprap and groynes may be placed in such way as to deliberately create a new channel shape. An example is the Kowhai River in Marlborough, which is being trained into a meandering course by placing lines of rock in the location desired for the new outer banks (Figure 16.14). This design was arrived at by a combination of regime theory, interpretation of existing channel patterns on aerial photographs, and experience.

River control works are designed to prevent bank erosion and channel migration, and to create a stable channel capable of carrying all bedload

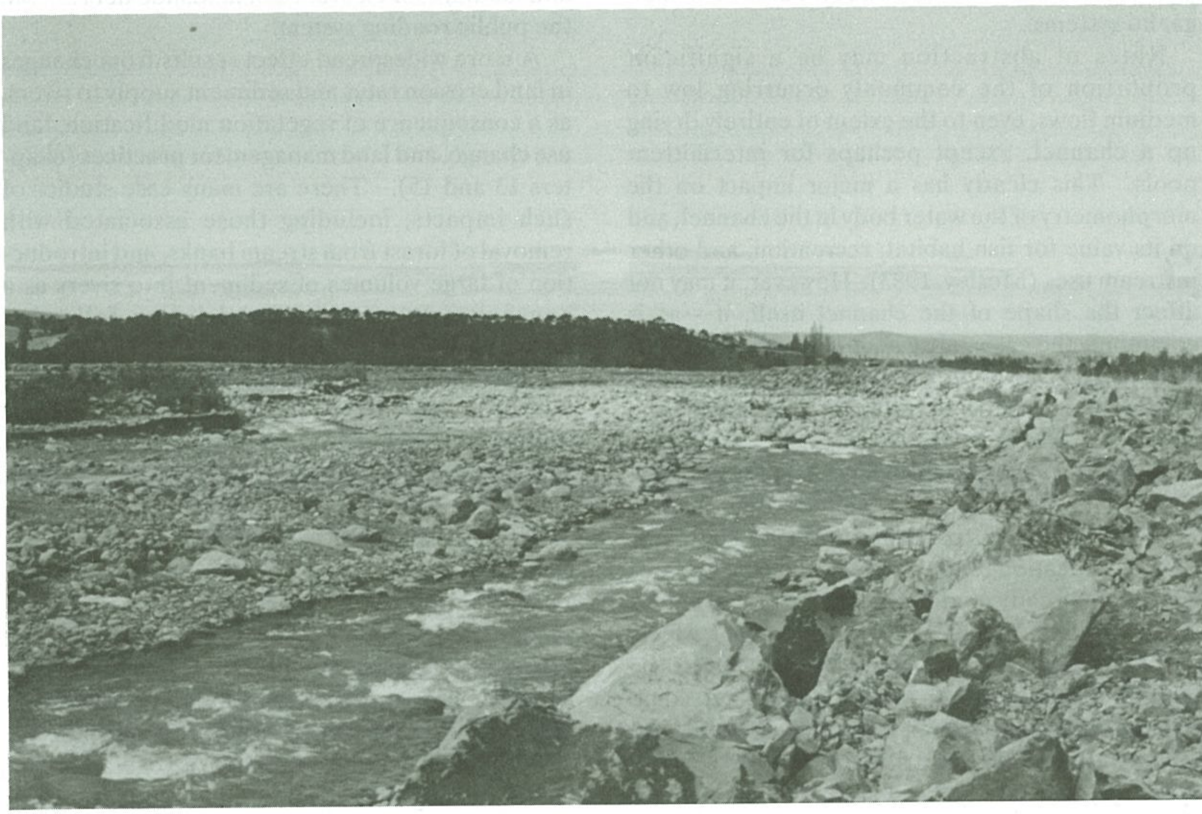


Figure 16.14 River training works along the Kowhai River near Kaikoura, designed to establish a meandering path.

through to the sea, without aggradation and loss of flow capacity. Regime theory (Blench, 1957) was developed overseas to aid in the design of stable canals, and has been applied to rivers in New Zealand, with rather less success. There have been several contributions to the development of design methods suitable for New Zealand's rivers (eg Griffiths, 1981), including the case where it is desired to convert a braided river to a single-thread channel (Griffiths, 1989).

The key requirement is selection of the size and shape of the channel which is able to convey the bedload sediment introduced at its upstream end, without aggradation or degradation. Carson and Griffiths (1987b) discuss some of the complications which make this difficult, particularly because of the complex inter-relationships between channel dimensions, shape, slope, flow conditions, and sediment transport, as well as the effects of fluctuating water discharge.

On most rivers in New Zealand existing works are largely effective and principally require maintenance and repair after major flood damage. Nevertheless, there are locations at which our lack of understanding of the interaction of water flow, sediment transport, and river morphology have hindered solution to a river control problem. One example is a reach of the North Ashburton, whose width has been reduced by intensive willow planting in an effort to increase flow depth and velocity, thus inducing sediment transport through the reach. Instead, however, the channel has continued to aggrade, to the extent that the bed now lies well above the surrounding farmland. Sediment transport per unit width increased, but the effective width across which sediment is transported decreased even more. This type of problem has been studied recently in a 1:50 scale model (Davies and Lee, 1988), and exemplifies the opportunities for future research into river morphology.

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17
Lakes

R H Spigel and A B Viner

Introduction

The apparent stillness of a lake conceals a complexity in physics, chemistry, biology and geology that lends the study of lakes its fascination and importance. Water movements in enclosed basins are weak compared with those of rivers and tides, but are of crucial importance for water quality, plant life and animal life in lakes.

Water in a river moves inexorably downstream, pulled downhill by gravity. Lake currents can flow in any direction, moving under the influence of wind, small density differences within the water itself, and (in large lakes) the earth's rotation. Variations in water density within a lake - *"density gradients"* - arise mainly from temperature differences, although concentration of dissolved salts or suspended sediments also can be important. Cold water is denser than warm water, salt water is denser than fresh water, and sediment-laden water is denser than clear water. Temperature variations in a lake are caused by solar heating during the day and cooling at night. One of the most striking differences between rivers and lakes is the degree to which small differences in density -in both horizontal and vertical directions, but mainly in the vertical - dominate circulation and mixing in a lake, and through these all of a lake's chemical and biological cycles.

Lakes serve a number of human needs. They provide storage for drinking water supplies and irrigation, ponding areas for flood control, potential energy for generation of electricity, sites for recreation, and paths for navigation and shipping. Lakes act as receiving waters for pollutants, and for sediments from both natural and human-induced erosion. Unlike rivers, which are constantly renewed by flows from upstream, lake waters are held within their basins for times measured in years, tens of years or longer. The much longer *"residence time"* for lakewater is another crucial difference between rivers and lakes. As a result lakes are vulnerable to the effects of human activities and development in their catchments - lake water quality and ecology can be easily damaged, and any deterioration is more difficult (and takes longer) to remedy than would be the case for rivers. Longer residence time also allows interactions to take place between water and sediment, and within the water column itself, that are not possible in a river. The resulting chemical and biological cycles are complex, often poorly understood and strongly influenced by the effects of density gradients on circulation and mixing. A lake integrates all of the human and natural impacts upon it. These include water, sediment and nutrient loads from rainfall. The response of a lake to the loadings on it depends strongly on the shape of its basin, and so we begin with a description of some of the processes involved in the formation and demise of lake basins, and of the resulting general characteristics of lake basin form or *"morphometry"*.

Fire and Ice: Lake Basin Origin and Evolution

Few places in the world can match the diversity of lake basin-forming processes that have occurred in New Zealand. The Lake District of Finland contains thousands of lakes that cover about half the total land area, but practically all are the result of glacial scouring and deposition during previous ice ages. Glacial action is also the major cause of lake basin formation in North America and Europe. In New Zealand lake basins have been formed by tectonic movements (faulting, subsidence, uplift) of the earth's crust, volcanic activity, glaciation, damming by landslides, river action, wind-blown scour and deposition, dissolution of limestone (karst lakes), coastal bar formation, and man-made dams. More than one cause may contribute to basin formation, and basins may undergo further gradual or sudden change after they are formed. Change can result in the eventual disappearance of a lake, by infilling due to sedimentation, or by drainage due to erosion of the geologic structure that forms a natural dam holding the lake waters in place. On the other hand, change can result in deepening or widening of an existing basin. Excellent reviews of the origin and development of lake basins in New Zealand are to be found in Lowe and Green (1987a), Irwin (1975a) and Pritchard (1989).

Lowe and Green point out that major lake types in New Zealand tend to be grouped into geographically distinct districts that coincide with the localised distribution of the major lake-forming geologic process. They have produced a map (redrawn here as Figure 17.1) that summarises the distribution and grouping of major lake types, as classified by their geologic origin.

Volcanic and Tectonic Basins

Few lakes in New Zealand are purely tectonic in origin, although tectonic movements associated with volcanic eruptions have played an important role in the formation of several volcanic lakes. New Zealand's volcanic lakes are located mainly in the central North Island in a relatively narrow band that extends from Mt. Ruapehu south of Lake Taupo northeast to White Island in the Bay of Plenty (Figure 17.1). This region contains all of New Zealand's presently active volcanoes, most of its active hydrothermal areas, and several individual volcanic complexes (Healy 1963; Cole and Nairn 1975; Nairn 1981).

Subsidence of the land surface following ejection of large volumes of molten rock and lava from subterranean magma chambers during volcanic eruptions produces large basins with a characteristic circular outline known as *"calderas"*. Lake Rotorua and Lake Taupo occupy their own calderas, although each has undergone stages of growth and contraction as a result of on-going volcanic activity, changes in drainage patterns and variations in rainfall since their formations approximately 140,000 and 300,000 to 350,000 years ago, respectively (Lowe and Green p.20, Timperley 1983 p.8). Lake Taupo is New Zealand's largest lake in terms of surface area (616 m^2), but not in terms of depth or volume - the South Island glacial lakes are deeper - and the many eruptions that formed the lake's different caldera basins have been particularly explosive. The 300,000 years B.P. eruption alone is thought to have produced more than 1200 km^3 of hot molten rock, ejected huge masses of volcanic ash and sulphur into the atmosphere, and was one of the largest eruptions in the world during the late Quaternary period (Froggatt *et al.* 1986).

The Haroharo caldera in the Okataina Volcanic Centre (Figure 17.2) is a large, complex geologic structure containing parts of the basins of Lakes Rotoiti, Okataina, Tarawera, Rotorua and Rotomahana. These basins developed their present forms within the past 8500 years or so, due to a combination of subsidence, damming by lava flows and shifting drainage patterns. The deep eastern basin of Lake Rotoiti, for example, lies on the rim of the caldera and originally filled when lava flows blocked eastward drainage of the caldera through the valley now occupied by the Tarawera River, forcing the Lake Rotorua - Lake Rotoiti system to drain northward by way of the Kaituna River. The western half of Lake Rotoiti occupies part of the old river channel formed by headward erosion before drainage to the east was

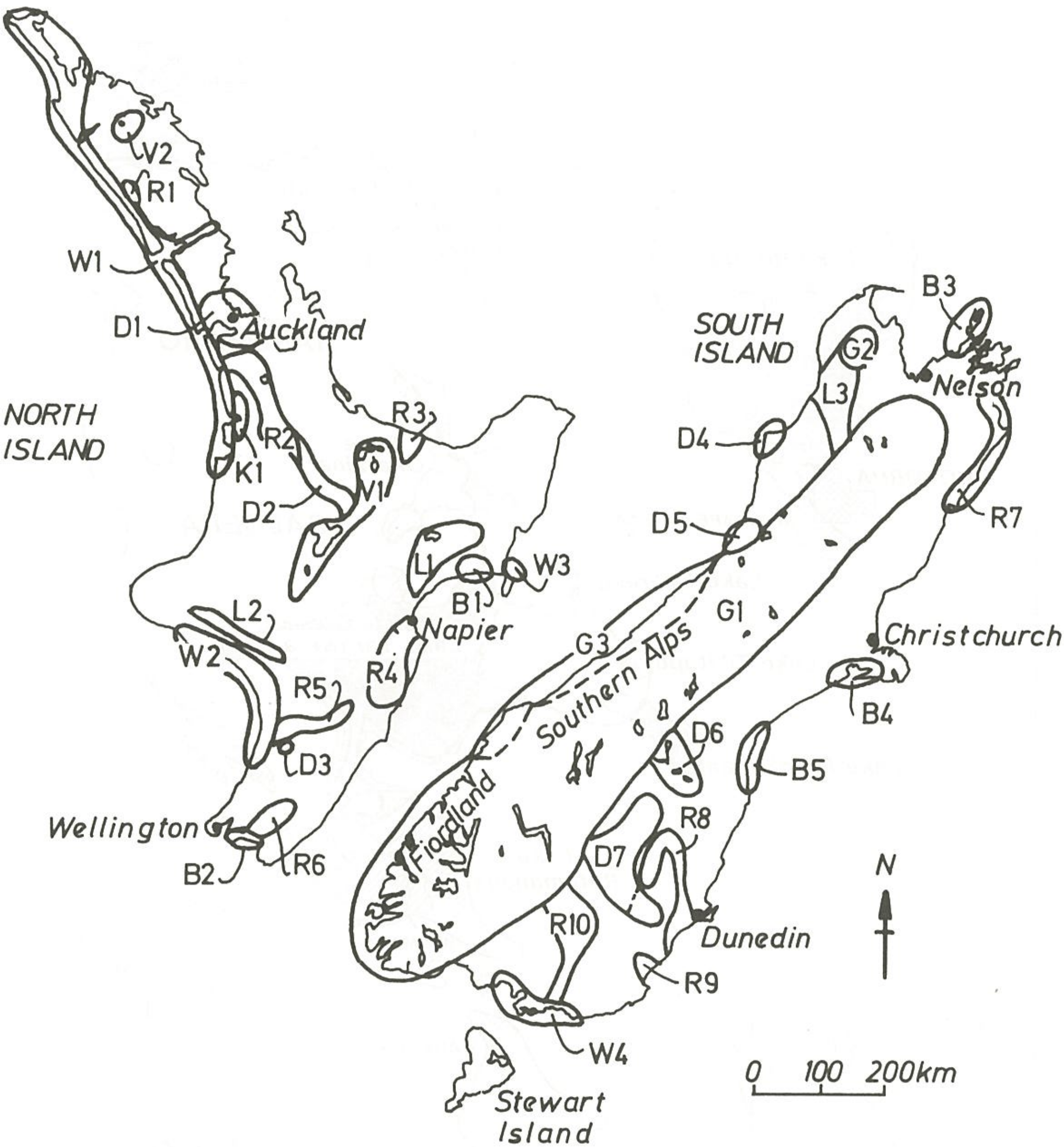

Figure 17.1 The distribution and grouping of major lake types, classified by geologic origin; redrawn from Lowe and Green (1987b, pp.460-1). Letters indicate lake type; symbols give the locality. B (barrier): 1 Hawke's Bay; 2 Palliser Bay, 3 D'Urville; 4 Christchurch; 5 Timaru. D (reservoirs): 1 Auckland water supply; 2 Waikato hydro-dams; 3 Manawatu; 4 and 5 Westland; 6 Waikato hydro-dams; 7 Clutha hydro-dams. G (glacial): 1 Alps; 2 Tasman glacier; 3 Westland. K (karst): 1 Waitomo. L (landslide): 1 Wairoa; 2 Wanganui; 3 Karamea. R (riverine): 1 Kaihu; 2 Waikato; 3 Bay of Plenty; 4 Hawke's Bay; 5 Manawatu; 6 Raumahanga; 7 Wairau; 8 Taieri; 9 Balclutha; 10 Mataura. V (volcanic): 1 Taupo-Rotorua; 2 Northland. W (sand-dune): 1 Northland, 2 Wanganui; 3 Gisborne; 4 Southland.

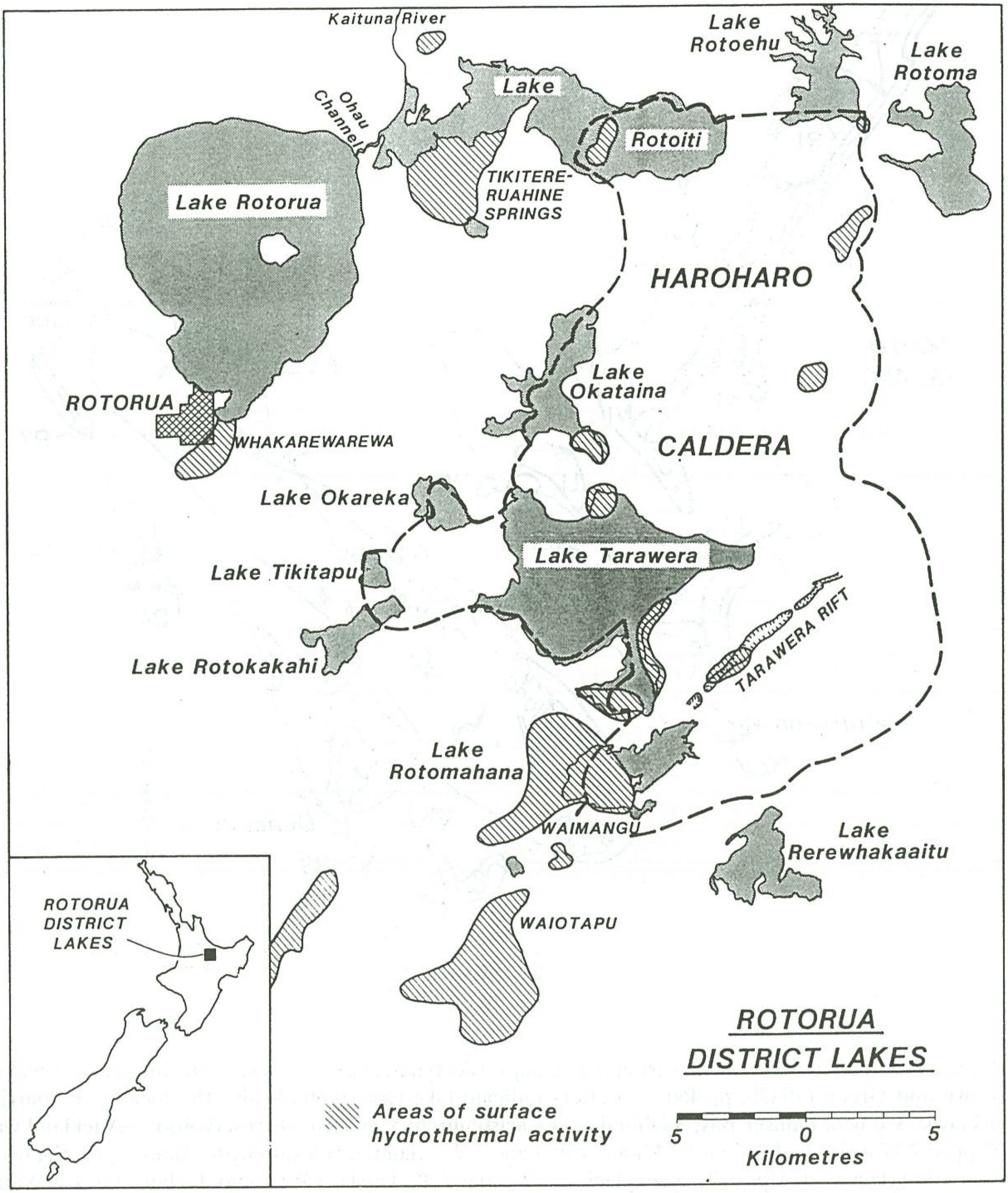

Figure 17.2 Rotorua district lakes, showing major areas of surface hydrothermal activity (based on Nairn 1981).

blocked. The most recent major eruption in the caldera occurred in 1886, resulted in 153 fatalities, and excavated a crater on a site originally occupied by two smaller lakes (Rotomahana and Rotomakariri) and the (then-) world-famous Pink and White Terraces (spectacular formations created by the precipitation of silica from hydrothermal waters flowing through the region). The crater took five years to fill and forms the basin of the present Lake Rotomahana. The lake has a maximum depth of 112 m and contains a geothermal heat source in its lake bed (Calhaem 1973, Nairn 1981). Geothermal heat flows have been mapped in the beds of nearby Lakes Rotoiti, Tarawera and Okataina by Calhaem.

Glacial Basins

Formation of lake basins by glacial action, although less spectacular than that of volcanoes, has nonetheless been of major importance in New Zealand. Out of a total of 776 New Zealand lakes for which Lowe and Green and Irwin (1975b) present statistics, 291 (38%) were formed by glacial action. While all of the volcanic lakes are in the North Island, all of the glacial lakes are in the South Island; of the total number (476) of South Island lakes, 61% are glacial in origin. This distribution reflects the extensive coverage of much of the Southern Alps by ice sheets during past ice ages, the most recent of which had its maximum approximately 18,000 years ago. As glaciers slowly advanced they eroded large, steep-sided *"overdeepened"* valleys (the bottoms of some being below sea level), stripping away soil and scouring bedrock. Hundreds of mountain tarns occupy hollowed *"cirques"* and depressions, from Fiordland through the Southern Alps to northwest Nelson (Figure 17.3a). Although small, some of these are well known and include the much-photographed Lake Quill above Sutherland Falls in Fiordland. The largest lakes of the South Island occupy long deep troughs that were gouged out by glaciers. These include Lakes Manapouri, Hauroko (New Zealand's deepest lake, maximum depth 462 m), Te Anau and Monowai in Southland; Wakatipu, Wanaka, Hawea and Ohau in Otago; Pukaki, Tekapo and Coleridge in Canterbury; and Rotoroa and Rotoiti (South Island) in Nelson.

As ice sheets melted and receded, they left behind deposits of rock, soil and sand known generally as glacial drift. Broad, thick deposits called moraines dammed rivers and plugged the ends of valleys already deepened by glacial scouring. Few of the large glacial lakes mentioned above are likely to be held completely by solid rock; for many glacial lakes, moraines probably contribute a small fraction of the total depth. Some lakes are enclosed completely by moraines, e.g. Lakes Brunner and Haupiri in Westland, but most glacial lakes are impounded by a combination of solid rock, moraine and contemporary alluvial outwash (Figure 17.3b,c,d). Lakes Pearson and Grasmere in Canterbury were impounded by talus fans or screes after the last glacial retreat approximately 10,000 years ago.

Other Formation Processes

Lakes associated with rivers are most common among the remaining types of basin. Small and shallow lakes commonly form in the lower floodplains of rivers in abandoned channels, cutoff meanders, and behind stopbanks. Basins can be created in coastal areas by beach and shoreline processes. Barrier-bar lakes are formed by longshore coastal sediment transport (e.g. Lakes Ellesmere and Forsyth, Canterbury). Lake Rotongaio, while not near the sea, may have formed as a barrier-bar lake as a result of longshore sand movement in much larger neighbouring Lake Taupo (Figure 17.5). Lakes formed by windblown deposits and sand dunes are common on the west coast of the North Island and range in age from a few hundred years to more than 50,000 years, with ages of 2,000-3,000 years being typical. Landslide-dammed lakes are less common; Lake Waikaremoana, the deepest (248 m maximum depth) and one of the least-modified of North Island lakes, was formed by a landslide across a deep gorge 2,000 years ago. It has a V-shaped cross-section and dendritic planform typical of a drowned river valley (and hence of most landslide lakes; but see Figure 17.3b).

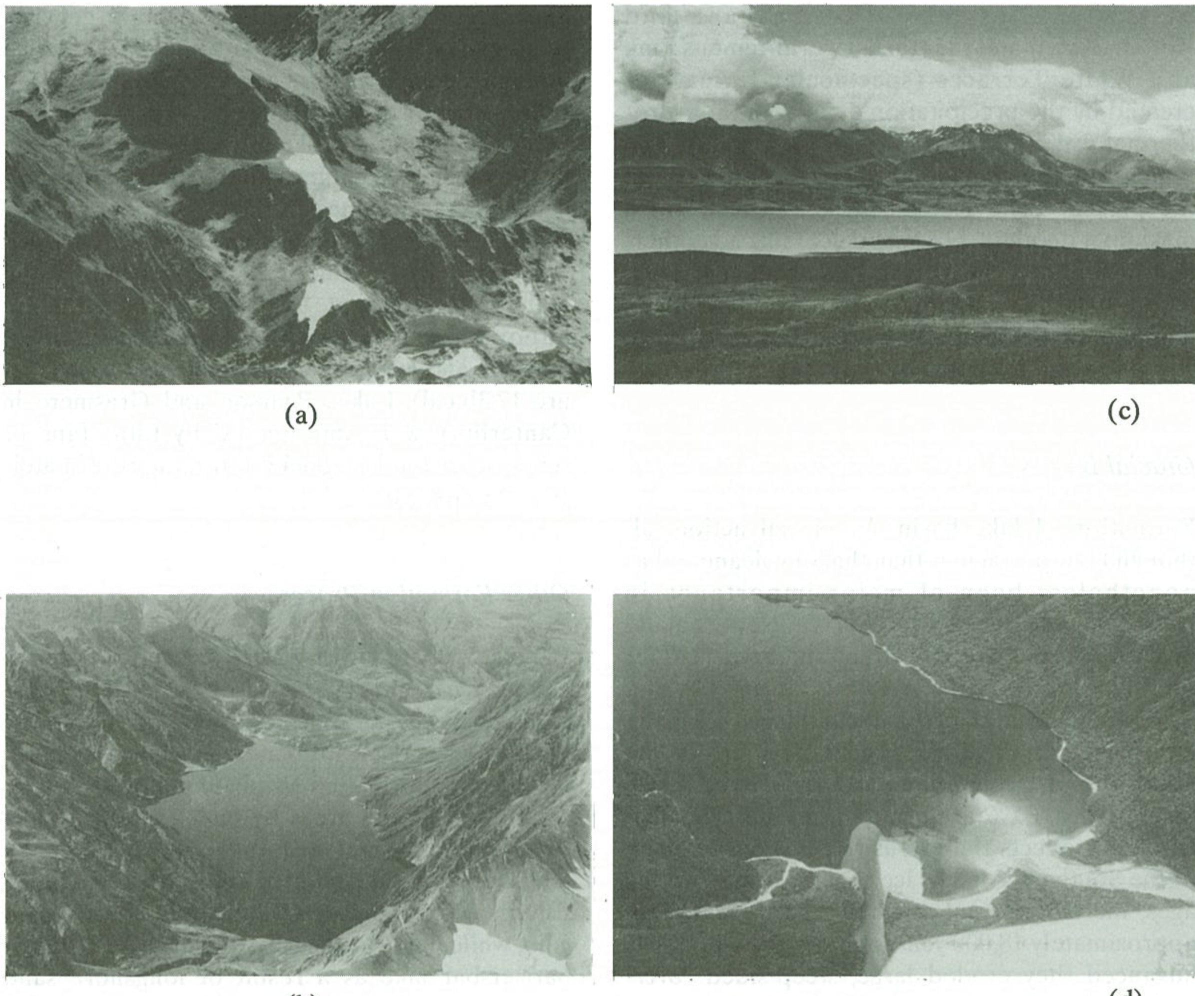

Figure 17.3 **(a)** Two unnamed cirque in the Southern Alps near the main divide above Lake Ohau. **(b)** Lochnagar in the catchment of the upper Shotover River, Otago, S. Island. The lake basin has been formed by glacial action but the lake is dammed by a landslide visible at the far end of the photo. **(c)** View across Lake Pukaki, a glacial basin, with moraines clearly visible on near and far shores. **(d)** Lake McKerrow, Fiordland, showing 3 inflowing streams: 2 branches of the Hollyford River and (on left) a small mountain stream. Sands transported by the Hollyford are white and are visible as a delta in shallow water. Most of the inflow discharge is in the left branch of the river, which appears turbid and is plunging to form an underflow. (All photos by M. MacSaveney.)

Reservoirs

Man-made lakes have been created in New Zealand mainly for water supply and power generation. The lakes that fill drowned river valleys behind dams tend to be long, narrow and sinuous in plan or may have a branching or dendritic planform if more than one river valley is dammed at a single site. In contrast to natural lakes, where maximum depths usually are found in a central part of the lake, water depth in reservoirs normally increases steadily in the downstream direction, with the deepest water near the dam. Water level fluctuations tend to be much larger in reservoirs than in natural lakes, and can leave large areas of bare soil around the lake perimeter subject to erosion by wind and rain. Most dams have offtake structures that allow water to be drawn from more than one level, in contrast to lakes where outflows come from surface water. All large dams must be built with adequate spillways to pass the largest possible flood expected to ever occur in the river behind the dam. Overtopping of a dam during a flood could cause the foundations of the dam to fail and the dam to collapse, causing catastrophic loss of life and property downstream.

The Auckland Regional Authority owns and operates several small water supply reservoirs in protected catchments in the Hunua and Waitakere ranges near Auckland. These lakes typically are a few kilometres long, a few hundred metres wide and less than 40 metres deep, many being located in hilly terrain and steep valleys. The largest artificial lakes in New Zealand are used for electric power generation. These include a chain of long, narrow lakes on the largest North Island river, the Waikato River draining Lake Taupo - Lakes Aratiatia, Ohakuri, Atiamuri, Whakamaru, Maraetei, Waipapa, Arapuni and Karapiro. In the South Island lakes have been formed for power generation high in the mountains of northwest Nelson (Cobb Reservoir), on the Waitaki River system in the MacKenzie Country of South Canterbury (Lakes Benmore, Aviemore and Waitaki) and on the Clutha River in Otago (Lakes Roxburgh and Dunstan). Several other natural lakes in the headwaters of the Clutha and Waitaki Rivers have had their major inlets and outlets modified so that water levels can be controlled for hydropower generation, and to connect the lakes into integrated hydropower networks. Lakes Tekapo, Pukaki and Ohau provide water for power generation in the Upper Waitaki power development scheme through a system of weirs, dams and canals. The Clutha River headwater lakes (Wanaka, Hawea, and Wakatipu) are not used directly for power generation, but their outflows are eventually impounded in Lake Dunstan behind the Clyde dam. Plans exist for further possible developments on the Clutha and Waitaki Rivers in the South Island and on the Motu and Mohaka Rivers in the North Island (Henriques, 1987). Several smaller hydro-lakes have been constructed and are maintained by local power boards and authorities throughout New Zealand.

Lake Bathymetry

The lake basins of almost all of New Zealand's largest lakes and many smaller ones have been surveyed. Each survey may involve hundreds of depth measurements or *"soundings"*, usually from a boat equipped with an echo-sounder to measure water depth and survey instruments to locate the position of the boat on the lake. Soundings are normally carried out along pre-set survey lines or transects (see Irwin 1972b and Hakanson 1981). The measurements can be summarized in a *"bathymetric chart"* that shows contours of depth in the lake basin; 120 of these charts have so far been prepared and published by the New Zealand Oceanographic Institute (NZOI) in the Lake Chart Series (Figure 17.4).

Lake Rotongaio (Figure 17.5a) is a small North Island lake separated by a sand-bar from Lake Taupo. A profile of the lake bed along a transect through the deepest sections of the lake is shown in Figure 17.5b; these figures will be used to illustrate some commonly used morphometric terms and depth-area-volume relationships. *"Maximum depth"* measured in Lake Rotongaio is 21 m. Areas within each depth contour, measured from the original NZOI Lake Chart, are given in Table 17.1; total lake surface area (zero depth in Table 17.1) is 34.84 hectares. Volumes are calculated by in-

Figure 17.4 Map showing location of lakes for which NZOI bathymetric charts have been prepared. (Note: L. Wahi = L. Waahi). (Reproduced from Holmes 1992; for further details contact the Publications Officer, National Institute of Water & Atmospheric Research, PO Box 14-901, Kilbirnie, Wellington.)

(a)

(b)

(c)

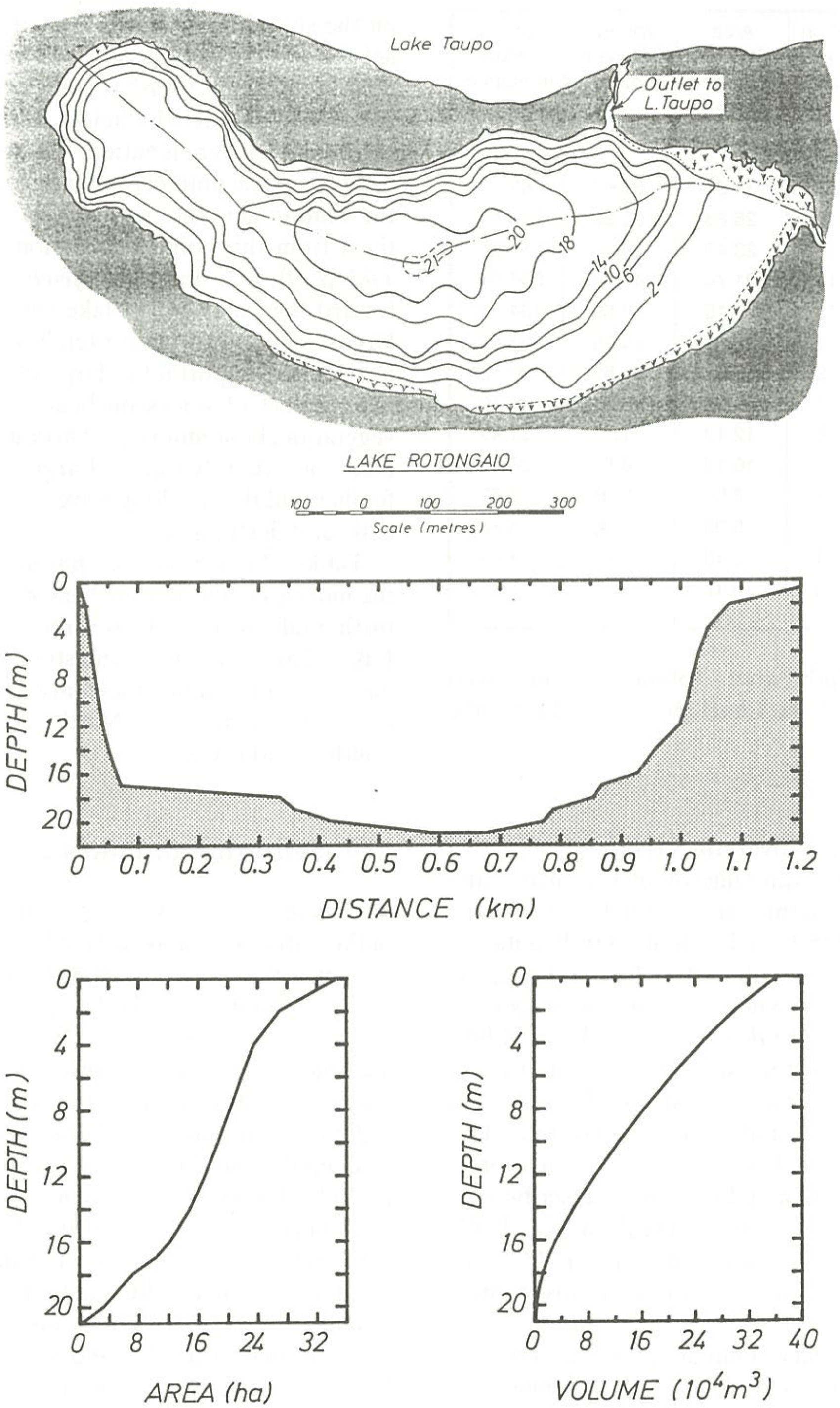

Figure 17.5 **(a)** Bathymetry for Lake Rotongaio, near Lake Taupo, Central North Island, based on Irwin (1978). Depth contours are in metres. **(b)** Profile of lake bed along the transect. **(c)** Hypsographic curve for Lake Rotongaio and volume below indicated depth contour, from Table 17.1.

Depth contour m	Elevation above bottom m	Area below indicated contour 10^4 m^2	Volume between contours 104 m^3	Volume below indicated contour 104 m^3
0	21	34.84	61.67	363.12
2	19	26.83	50.25	301.45
4	17	23.42	45.16	251.20
6	15	21.74	41.89	206.04
8	13	20.15	38.69	164.15
10	11	18.54	35.36	125.46
12	9	16.82	31.65	90.10
14	7	14.83	26.96	58.45
16	5	12.13	11.13	31.49
17	4	10.13	8.59	20.36
18	3	7.04	6.05	11.77
19	2	5.05	4.08	5.72
20	1	3.10	1.64	1.64
21	0	0.17	-	

Table 17.1 Depth - area - volume data measured from the NZOI Lake Chart (Irwin 1978) for Lake Rotongaio

tegration of area over the depth. Alternative methods for approximating volumes from area and depth measurements are given by Hakanson (1981). Area-depth and volume-depth data for Lake Rotongaio are plotted in Figure 17.5c; the area-depth plot is known as the *"hypsographic curve"*. *"Average lake depth"* is given by total lake volume divided by the lake surface area; it is the depth of a basin that has the same surface area and volume as the lake, but with a flat bottom and a constant area equal to the lake's surface area. Many other terms may be used to describe lake basin shape and outline (Hakanson, 1981; Hutchinson 1957). *"Maximum lake length"* is the length of a line connecting the two most remote points on the shoreline; it may not cross the shoreline of the lake boundary, but may cross islands and may be curved. In Lake Rotongaio the maximum length is approximated by the length of the transect line shown in Figure 17.5, 1200 m. The *"maximum effective length"* is the length of a straight line connecting the two most distant points on the shoreline over which wind and waves may act without interruptions from land and islands. Finally, *"effective fetch"* is sometimes used in formulae that predict the height or energy of wind-waves at a given location. Effective fetch is a measure of the uninterrupted distance available to the wind to generate waves over a range of directions from the point in question (see Hakanson 1981 p.29). For *"maximum fetch"* the direction is toward the point on the lake shore that gives the largest value for effective fetch. Effective fetch is particularly important in large lakes to assess the likely effect of waves on beach erosion, aquatic vegetation, boat moorings, harbours, breakwaters, piers or other features. Large lakes have long fetches and the resulting waves can be large, energetic and destructive.

Table 17.2 summarises parameters related to the morphometry of some New Zealand lakes. For further information, the Inventory of New Zealand Lakes, Parts I and II (Livingston *et al.* 1986), summarises information on several aspects of lakes and their catchments for 81 North Island lakes and 84 South Island lakes.

Sedimentation in Lakes and Reservoirs

Lakes and reservoirs act as sediment traps. Most of the sediment that is carried by rivers into a lake, or that settles to the bottom from the death and decay of plankton in the lake, remains within the confines of the lake basin. Over time, then, lakes are bound to become shallower and shallower until they eventually disappear. In some lakes, however, sedimentation rates may be so slow and the lakes so deep that infilling appears negligible over long periods of time. Also, tectonic or volcanic processes, climate change or shifting drainage patterns may increase basin depth or water volume at a greater rate than sedimentation decreases them. Finally, sedimentation does not occur uniformly over the bed of a lake, and sediments that have been deposited are sometimes resuspended, transported to different locations and redeposited.

Lowe and Green give estimates of average sedimentation rates ranging from less than 1 mm/yr to almost 200 mm/yr. North Island vol-

Table 17.2 Characteristics of some New Zealand Lakes

Lake North Island	Surface Area km^2	Average Depth m	Maximum Depth m	Maximum Length km	Maximum Width km	Altitude m asl	Type*
Taupo	622.63	97	162.8	40.5	29.5	357	V
Wairarapa	79.84	1.5	2.5	18.2	9.6	0.1	R
Rotorua	80	11.0	44.8	12.1	9.7	280	V
Waikaremoana	55.74		248	16.0	11.1	585	L
Tarawera	41.02	50	87.5	11.4	9.0	299	V
Rotoiti	34.35	33.04	122	15.0	3.6	279	V
Rotaira	15.32	8.9	14.6	6.3	3.6	564	V
Rotoma	11.16	36.9	83.0	5.2	4.7	313	V
Okataina	10.8	39.4	78.5	6.2	5.0	311	V
Rotoehu	8.11	8.16	13.5	4.6	4.0	295	V
Rotomahana	7.95	60	125	6.2	2.8	335	V
Rerewhakaitu	7.47	7.0	15.8	3.8	3.7	438	V
Rotokakahi (Green Lake)	4.48	17.5	32.0	4.3	1.7	394	V
Okareka	3.46	20.0	33.5	2.8	1.9	355	V
Pupuke	1.63	34.0	55.0	1.3	1.0	0.3	W
Tikitapu (Blue Lake)	1.40	18.0	27.5	1.6	1.3	417.8	V
Rotongaio	0.35	10.42	21.0	1.2	0.4	357	B
Okaro	0.28	12.1	18.0	0.7	0.6	423	V
Artificial Lakes							
Ohakuri	7.21		37.5	7.5	2.0	289	D
Whakamaru	7.2		38	15.0	1.0	229	D
Waipapa	5.5		16.5	1.6	0.3	129.5	D
Karapiro	5.37		30.5	11	0.9	55	D
Maraetei	5.0	17	61	7.2	1.2	183	D
Arapuni	4.95	53	64	6.5	0.9	113	D
Otamangakau	1.8	12.4	3.6	0.7		665	D
Atamikmuri	1.68	11.6	28.5	4.7	0.4	254	D
Aratiatia	0.34		5.0	2.2		360	D

Lake South Island	Surface Area km^2	Average Depth m	Maximum Depth m	Maximum Length km	Maximum Width km	Altitude m asl	Type*
Te Anau	347.50	132	417	60	28.6	203	G
Wakatipu	289.17	210	380	75.2	6.2	310	G
Ellesmere	181.75		2.1	26.3	12.9	0.30	B
Wanaka	180.1		311	45.5	11.6	277	G
Manapouri	143.33		444	28.3	11.5	179	G
Hawea	137.60		384	41.9	10.4	347	G
Pukaki	98.90		70.0	22.9	8	494	G
Tekapo	86.80		120	25.2	5.9	708	G
Hauroko	68.30		462	33.7	7.8	157	G
Poteriteri	42.5			27.2	3.0	23	G
Ohau	53.85	74.5	129	16.8	5.1	517	G
Brunner	36.10		109	9.4	6.8	86	G
Coleridge	32.90		17.8	200	3.4	507	G
Monowai	32.5		161	20.6	2.5	206	G
Rotoroa	21.40	96.5	152	14.4	2.9	444	G
McKerrow	18.3		121.3	15.3	2.2	3	G/B
Kaniere	14.65		197	8.6	2.6	132	G
Sumner	11.80		134.5	9.8	2.1	521	G
Rotoiti	9.20	49.2	82.0	8.5	2.6	609	G
Alexandrina	5.80	13.4	30	6.9	1.3	730	G
Hayes	2.03	18.7	35	3.1	1.1	329	G
Taylor	1.85		40.5	3.1	0.9	582	G
Pearson	1.79	7.3	17.0	3.7	1.2	607	G
Lyndon	1.11		28	3.9	0.5	15.55	G
Grasmere	0.67	7.8	15	1.5	0.6	583	G/R
Quill	0.53			1.1	0.1	985	G
Matheson	0.3		12	0.7	0.2	110	R
Artificial Lakes							
Benmore	68.6		120	26.1	6.1	360	D
Dunstan	25	14	55				D
Aviemore	24.8		62	17.6	4.5	271	D
Mahinerangi	18.6	6.2	31.2	21.5	3.4	391	D
Waitaki	6.2	6.4	21.4	7.2	2.3	230	D
Roxburgh	5.9	9.9	42.7	28.5	0.67	133	D
Cobb Reservoir	1.55		20	5.9	0.3	808	D

Notes:

(1) Information from Livingston **et al.** 1986, Lowe and Green 1987b, Winter 1963, Thompson 1976, Jowett and Thompson 1977, McKenzie 1987, Works Consultancy - Power Engineering and other sources.

(2) Average depth = volume/surface area.

(3) Catchment area includes lake surface area.

* V = Volcanic, R = Riverine, L = Landslide, W = Wind-blown sand (dune), B = Barrier, D = Dam, G = Glacial.

canic lakes with small river inflows have low rates, on the order of 1 mm/yr; sedimentation in these lakes is dominated by settling of dead plankton. South Island lakes fed by glacial streams that carry a high suspended sediment load (e.g., Pukaki and Tekapo) have sedimentation rates on the order of 10-15 mm/yr. The highest sedimentation rates (100-175 mm/yr) apply to reservoirs on rivers that flow through erodible country and transport large amounts of sediment, especially in flood.

Processes governing sediment inflow, deposition, and redistribution in lakes are illustrated in Figure 17.6. When a river enters a lake the river's momentum, and hence ability to transport sediment, decreases abruptly and the heavier sediment moving either along the bed or in suspension is deposited to form a delta at the river mouth. What happens subsequently to the river inflow and the remaining sediment it is carrying depends largely on the density difference between the river water and the lake water. Generally this density difference is controlled by temperature, but suspended sediment concentrations can also be important, especially if these are high. If the river is less dense (e.g. warmer) than the lake, the river flow will lift off from the lake bed and spread over the surface of the lake as an *"overflow"*. If the river density is the same as that of the lakewater, the river will mix with the lake over the entire water column, until its momentum is completely dissipated. If the river is denser than the lakewater (because of either temperature, suspended sediment load or a combination of the two), then the river will plunge under the lakewater and flow

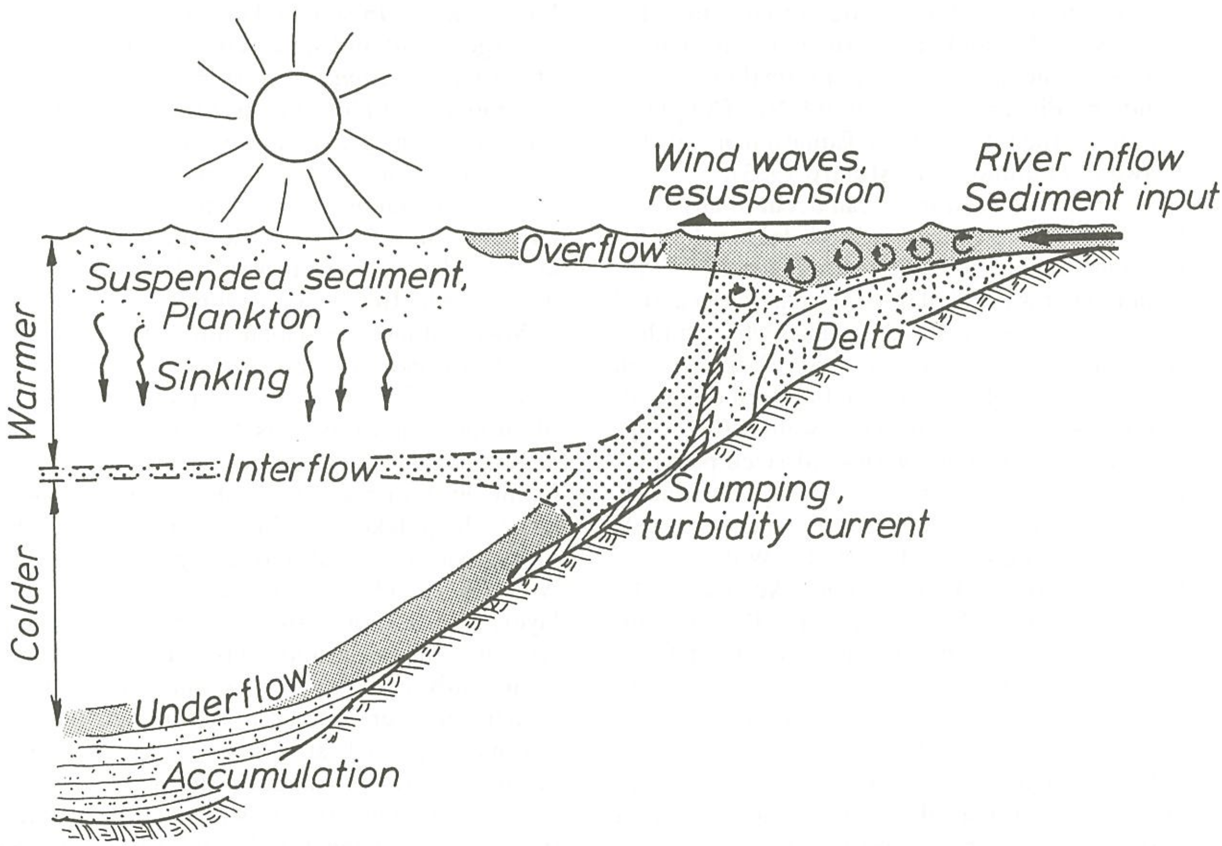

Figure 17.6 Sedimentation processes in lakes. Also shown are the effects of density differences (whether caused by temperature, salinity, or sediment) between inflow and lake water in causing underflows, interflows, or overflows.

down the bed of the lake as a submerged river or *"underflow"*. Considerable mixing of river and lakewater often accompanies the plunge, with somewhat less mixing as the river continues down the steepest slope of the lake-bed. If the underflow reaches a depth where its density matches that of the surrounding lake water, then its downhill momentum will be halted and it will lift off from the lake bed and form an *"interflow"* or *"intrusion"* at depth, spreading out horizontally along a surface of constant density in the water. The different inflows and associated sedimentation regimes are summarized in Figure 17.6.

The dependence of lake water density on temperature, salinity and pressure can be calculated using a formula presented by Chen and Millero (1977). Although salinity and pressure effects on water density are unimportant for most New Zealand lakes, further information can be found in oceanography textbooks (e.g. Pond and Pickard 1983) and in UNESCO publications on the International Equation of state for Seawater (Fofonoff and Millard 1983).

Deposition of sediment can occur at any stage for any inflow pattern, with the larger, heavier particles tending to be deposited closer to shore in shallower water and the finer particles carried further offshore. However, in lakes where underflows are common and reach relatively high speeds, coarser fractions can be carried considerable distances. In some lakes well-defined submarine channels, lined with sand-sized particles in contrast to the finer silts and clays elsewhere, mark the paths of these flows. Such channels have been described in Lake Wakatipu by Brodie and Irwin (1970) and in Lake Wanaka by Irwin (1980). In Lake Wakatipu Pickrill and Irwin (1982) have found that winter river flows are cold enough to underflow continuously. In summer, rivers are warmer and either underflow or overflow, but floods can enter as sediment-laden underflows, reaching speeds of 0.2 m/s (a high velocity for a lake current!) and carrying particulate matter over 60 km downslope into the deepest park of the lake. Turbulence diminishes rapidly after a river inflow lifts off from the lake bed to form an intrusion. Only clay-sized particles will remain in suspension for any distance in such an intrusion, and these will settle slowly. Overflows are subject to wind mixing, and the resulting turbulence may help keep sediment in suspension, but generally this is less efficient than turbulence induced by the lake bed for an underflow.

Sediments deposited on the downward sloping sides of a lake-bed may be transported to deeper parts of the lakes. Earthquakes, wave action (in shallower water), over-steepening of sediment surfaces leading to slumping, even alternate drying and then sudden flooding of exposed sediments - all can trigger sediment movement downslope. Offshore resuspension and transport of sediment from a submarine delta formed by a glacially fed inflow has been described by Pickrill and Irwin (1983) for Lake Tekapo. In some lakes the movements can be large, rapid and spectacular, taking the form of *"turbidity currents"* - very muddy underflows that move with high speeds. Once set in motion these flows may accelerate - because of their high speed, the flows entrain more sediment, thereby becoming heavier, moving even faster and entraining still more sediment. This effect may apply to underflows in Lake Wakatipu described by Brodie and Irwin (1970).

Most sediment resuspension and redeposition mechanisms move sediment gradually downslope into the deeper parts of lakes, a phenomenon known as *"sediment focusing"*. Sediment deposits in lakes can be tens of metres in thickness, or even hundreds of metres in older, large, deep lakes. Sediment cores from lakes often exhibit a distinctive layered structure. Average particle size tends to vary from layer to layer, alternating between larger and smaller sizes in a quasi-periodic fashion. The layering occurs only in lakes where sediment is carried in suspension over wide areas before settling. The alternating particle sizes reflect seasonal, annual or longer-period changes in runoff and sediment load in inflowing rivers, with coarse fractions transported during floods and finer fractions carried by low flows. By examining the chemistry, layering, and fossil content (often of pollen) of cores extracted from deep, undisturbed sediments, scientists can piece together clues about changes in past climates, changes in plant and animal communities, and patterns of human settlement. While such *"paleolimnological"* studies are more common in North America, European and Japanese lakes, studies have been done in New Zealand to examine the changes in runoff patterns and the effects of human settlement in the catchments of Lake Taupo (Rawlence and Reay 1976) and Lake Rotorua (Rawlence 1984).

Accumulation of sediments can be a major problem in lakes and reservoirs used for water supply or power generation. Infilling of such lakes leads to loss of valuable storage capacity. Small reservoirs can be sometimes drawn down sufficiently to excavate accumulated sediments, but in large reservoirs this is not practical. In Lake Roxburgh on the Clutha River, bed profile surveys made between 1956 and 1989 (Thompson 1976; Works Consultancy Services - Power Engineering, pers. comm.), showed the reservoir was filling at the rate of 1.8 million cubic metres (bulk sediment volume) per year. The volume of the lake decreased from 117×10^6 m^3 in 1956 to 58.6×10^6 m^3 in 1974, a decrease of 50% of the original lake volume in 33 years; averaging this volume over the entire lake area implies an average sedimentation rate of 300 mm/yr. At this rate sediment would have completely filled the lake in approximately 66 years. Yet the expected life of the reservoir was 300 years, based on the little data available in 1905 (pers. comm. - A.J. Sutherland). Clyde Dam, now completed upstream, will prolong the life of Lake Roxburgh.

Not all reservoirs experience such problems. Extensive seismic profiles made by Pickrill *et al.* (1984) of sediments in Lake Arapuni on the Waikato River indicated an annual sedimentation rate of 4-11 mm/yr. Coarse sediments were found only in the upper reaches of this long, narrow lake, with much thinner layers of mud deposited in the middle and lower reaches. Arapuni was the first dam built on the Waikato River, and Pickrill *et al.* concluded that the lake probably experienced relatively high sedimentation rates only from 1929 until 1952, when upstream dams were commissioned on the Waikato. The only river-borne sediments to reach the lake now are fine particles in suspension in the Waikato River.

While lake sediments are important because of their accumulation, of equal importance are the chemical reactions that take place between sediments and the water. In the central part of most lakes there is a steady downward rain of not only suspended sediments washed in by rivers and rain, but also of dead plankton. (*"Plankton"* are the floating microscopic plants and animals found in lakes, rivers and oceans.) As they sink, the dead plankton are colonised by bacteria and fungi that break down complex organic compounds into simpler inorganic compounds of carbon, nitrogen, phosphorus, hydrogen and oxygen, as well as of *"trace elements"* such as iron, manganese, silica, and sulphur. This *"mineralization"* is carried out most efficiently when there are adequate supplies of dissolved oxygen in the water, and is necessary before the carbon, nutrients and trace elements can be utilized for further growth by living plankton. In very deep lakes, and where dissolved oxygen is present throughout the depth, mineralization may be completed before particles reach the bottom. But often mineralization is only partially completed before particles settle and are covered by further deposits, thereby slowly building up a rich layer of organic ooze on the lake bed. These organic muds and the deeper lake waters (the *"hypolimnion"*) form a reservoir of carbon, nutrients and trace elements that are available for recycling within the lake. The complex chemical pathways that make up the cycles of oxygen, carbon, nitrogen, phosphorous and other elements are intimately related to chemical reactions and microbial activities that occur at the sediment-water interface. It is here that bacterial populations are highest (Wetzel 1975 p.593) and that compounds essential for plant and animal growth can be either released in readily available dissolved form, or precipitated and locked away in inaccessible form, or, in the case of nitrogen, even lost to the atmosphere as a gas (*"denitrification"*). Moreover, all of the cycles are interdependent (most importantly with availability of oxygen), and also depend on circulation and mixing (and hence density stratification) within the lake. These cycles ultimately control the plankton productivity, water quality, fishery and aesthetics of a lake. Further information may be found in textbooks on limnology such as Wetzel (1975). More information on the physical processes involved in sedimentation in lakes, including sampling techniques and particle-size classifications, can be found in Hakanson and Jansson (1983).

Water Balance in Lakes

Water balance calculations for lakes have many applications. They are used, for example, to predict the change in lake level due to withdrawals of water for power generation, irrigation, or water supply; to trace the progress of a flood wave as it travels through a lake or reservoir; or to determine the storage capacity that a reservoir would need in order to smooth out variable river inflows and provide a steady supply of water for human consumption.

All lakes receive water from precipitation directly on their surface and lose water by evaporation. Precipitation, evaporation and other sources and losses of water for most lakes are illustrated in Figure 17.7. If the flows into a lake exceed the outflows, then the lake level will rise, and the volume of water in the lake will increase (rainfall and evaporation being included as flows in and out). Similarly, if outflows exceed inflows, water volume will decrease. A water balance is simply an expression of conservation of volume, equating rate of change in volume to inflows minus outflows:

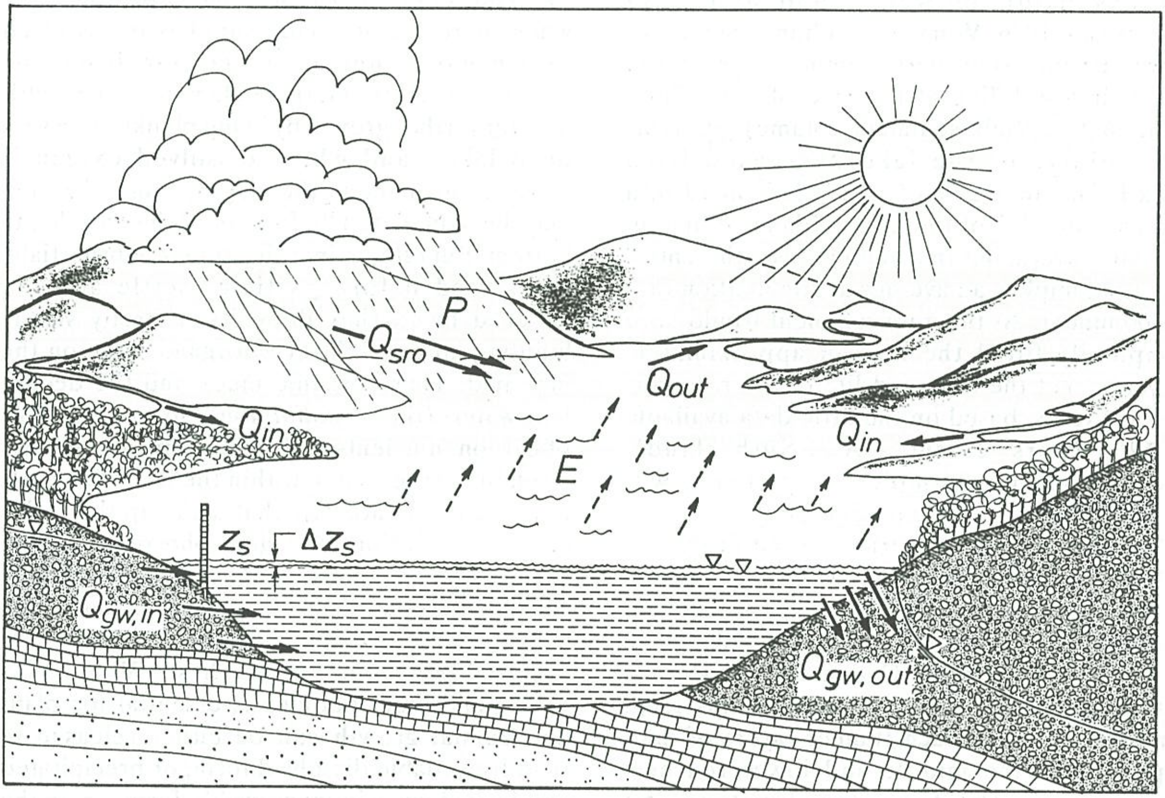

Figure 17.7 Water balance components: river inflows (Q_{in}); storm runoff (other than in river inflows) (Q_{sro}); river outflow (Q_{out}); groundwater inflow and outflow ($Q_{gw,in}$ and $Q_{gw,out}$); and rainfall on, and evaporation from, the lake surface (P and E). Changes in storage are computed from changes in lake level (Δz_s).

$$dV/dt = \Sigma Q_{in} - \Sigma Q_{out} + Q_{sro} + Q_{gw,in} - Q_{gw,out} + (P - E)A_s \quad (17.1)$$

where V = volume of water in the lake at time t; dV/dt = rate of change of V; ΣQ_{in} = sum of all river inflows; ΣQ_{out} = sum of all surface water outflows, including natural river outflow and artificial withdrawals through canals, reservoir offtake structures or pipes; Q_{sro} = surface-water runoff during storms from the lake catchment that has not been included in ΣQ_{in}; $Q_{gw,in} - Q_{gw,out}$ = net groundwater inflow; P = precipitation on the lake surface; E = evaporation from the lake surface; and A_s = surface area of the lake. The relative importance of the various terms in the water balance will change from lake to lake and even within one lake. The long-term water balance of Lake Rotoiti (N. Island) has been calculated by Spigel (1989), using monthly data from 1957 to 1984. Lake Rotoiti receives its only major river inflow from Lake Rotorua via the Ohau Channel which enters the lake at its western end. Outflow is to the Kaituna River, the outlet being nearly adjacent to the Ohau Channel inlet (Figure 17.2 and 17.8). Most of the surrounding volcanic soils are highly porous so surface streams are small and groundwater flows are important. Groundwater seeps into the lake on its southern and eastern shores, and the lake probably loses groundwater to seepage from its northern boundary. When averaged over a long period the changes in lake storage were not significant, with the balance being between inflows and outflows (the right-hand-side of Eq. 17.1): Ohau Channel inflow = 18.04 m^3/s; Kaituna River outflow = 22.82 m^3/s, P = 2010 mm/yr (2.19 m^3/s for A_s = 34.35 km^2), E = 880 mm/yr (0.96 m^3/s), and $Q_{gw,in} - Q_{gw,out} + Q_{sro}$ + inflow from small surface springs and streams + errors = 3.55 m^3/s. These latter terms were grouped together as a residual in the balance as they could not be measured directly. Errors for the overall balance were estimated as

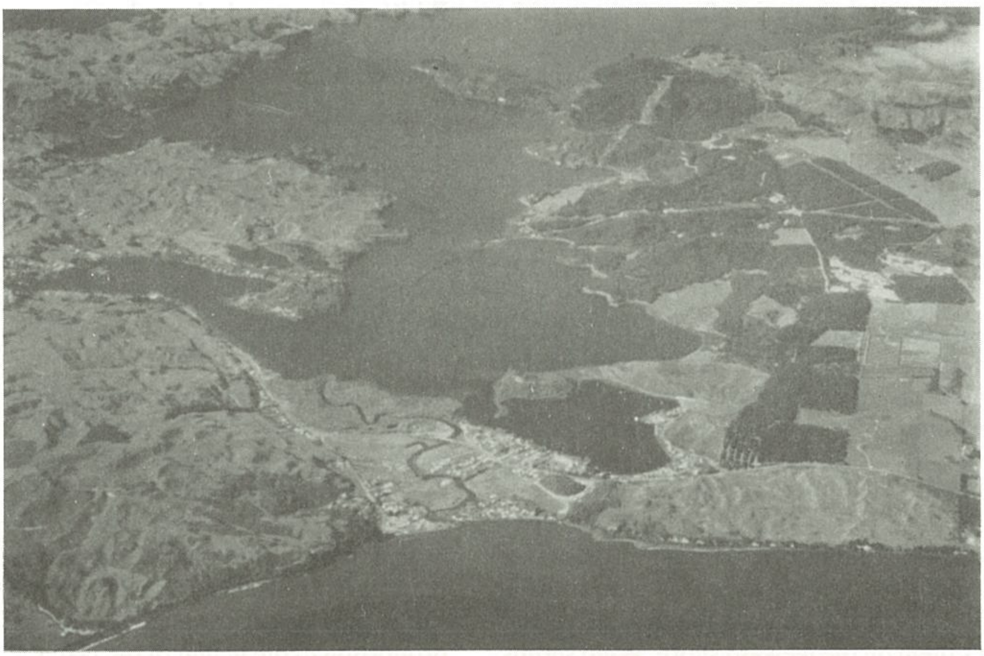

Figure 17.8 The western basin of Lake Rotoiti (N. Island), with Lake Rotorua and the Ohau Channel in the foreground, and the deeper central basin to the east in the background. The northern arm of the western basin (to the left in the photograph) leads to the lake outlet to the Kaituna River. (Photo by M. Gibbs.)

±2.5 m^3/s, somewhat larger than the variation found in the residual term (standard deviation of the residual = 1.85 m^3/s). The relative constancy of the residual, compared with the large variations in the other terms, and taking into account likely magnitudes of small surface streams, showed that there is almost always net groundwater inflow to the lake.

Change in Volume: Lake Levels and Seiches

Some of the longest and most reliable hydrological time-series in New Zealand are of lake levels. Measurements for Lake Taupo began in 1905, for Lake Rotoiti (N. Island) in 1906, for Lake Wakatipu in 1923, and Lake Wanaka in 1929; similar record lengths exist for some other lakes (Walter 1990). Many of the early records are from staff gauges read daily or less frequently. Today lake levels are recorded automatically at 15-minute intervals in many New Zealand lakes (Walter 1990). Several measurement sites are at or near lake outlets, since there is often a very stable relationship between lake level and river outflow. These lake level records serve a dual purpose, providing information on both storage volumes and river outflows.

Changes in lake storage are related to changes in lake level by:

$$dV/dt = A_s\, dz_s /dt \qquad (17.2)$$

where A_s = lake surface area and z_s = the lake-wide average water level. The elevation measured at a single point by a water-level recorder may not be an accurate measure of z_s, however, as the surface of a lake is seldom horizontal. In addition to the familiar wind-waves and swell that are present on all but the calmest days, winds also generate basin-scale tilts or *"setups"* of the water surface. The drag exerted by a steady wind on a lake raises water levels on the downwind shore and depresses them on the leeward shore. Wind setups range from a few centimetres to 20 cm in deep lakes such as Taupo, to well over a metre in long, shallow lakes such as Ellesmere. Even a few centimetres, when multiplied by the surface area of a large lake to calculate a volume (Eq. 17.2) can result in a significant error in the water balance if a setup is interpreted as a lake-wide change in water level. When the wind stops the water surface eventually returns to a horizontal, equilibrium position. However, the water surface usually overshoots the horizontal, tilts in the opposite direction, returns, and continues to rock about the horizontal in a see-saw fashion until friction between the water and lake-bed dissipates all of the energy in the water movements. These basin-scale standing waves, known as *"seiches"*, are observable at the shoreline as a gradual rising and falling of water with intervals of 20-40 minutes typically separating successive high-water levels. Seiches with a period of 36 minutes have been recorded in Lake Taupo, 39 minutes in Lake Wanaka (Thompson and Ibbitt 1978) and 52 minutes in Lake Wakatipu (Heath 1975); shorter periods would be expected in smaller lakes. These oscillations can introduce errors in the water balance unless the water level record is averaged over a sufficiently long period of time. Thompson and Ibbitt recommend three hours, and point out that too much averaging can result in a loss of genuine information about changes in lake level. Setups and seiches can also be generated by changes in atmospheric pressure (such as those accompanying storm fronts), flood inflows, landslides into lakes, or seismic disturbances in the lake bed.

River Inflows and Outflows

Lake basins occupy depressions in the landscape and therefore tend to collect runoff from one or more streams or rivers. Lakes commonly have several river inflows but only one river outflow, and some lakes have no river outlets. A river outlet occurs where the lake level rises above the lowest point on the rim of its basin. The amount of outflow will vary in direct (though not necessarily linear) proportion to the height of the lake's water surface above the outlet. This tends to stabilize lake levels and reduce the range over which they vary compared with lakes with no surface outlet. Lakes with a surface outlet are said to occupy *"open basins"*, while those with no river outflow occupy *"closed* *basins"*. The term *"closed basin"* is sometimes reserved for lakes that lose water only by evaporation and have neither surface nor groundwater outflow. Lakes with groundwater outflow are them termed *"seepage lakes"* (Hutchinson 1957, p.231).

Groundwater

Groundwater contributions to the water balance are very difficult to assess, and almost impossible to measure directly. Groundwater may enter a lake along a shore where the water table is above the level of the lake, and leave the lake through the bed where the water table is below the level of the lake (Figure 17.7). Groundwater can also enter from submerged springs through an otherwise impermeable lake bed if the piezometric surface of the groundwater is above the lake's water-surface. Lake beds tend to be relatively impermeable where they are covered with clays and organic material. However, the direction and magnitude of groundwater flow depends not only on permeability but on differences in hydraulic or piezometric head as well. Groundwater inflows and outflows can be highly variable because of changes in water-table elevations due, for example, to recharge by rainfall or withdrawals by wells, and because of variability in topography and geology. A method for measuring groundwater inflows at specific locations on a lake bed was described by John and Lock (1977) and Lock and John (1978), who applied the technique in the near-shore regions of Lake Rotorua and Lake Taupo. Such techniques, while extremely useful for water quality studies and for proving conclusively the direction and magnitude of local groundwater flows, cannot hope to give lakewide results for a water balance. Often the only way to estimate net groundwater inflow is to treat $Q_{gw,in} - Q_{gw,out}$ as a residual or an unknown in Eq. 17.1 and try to measure all the remaining terms directly. In many water balance calculations net groundwater inflow is simply lumped with an error term because it is impossible to measure or calculate directly.

Numerical groundwater models can be used to calculate lake - groundwater inflows and outflows. Hunt (1976) used a mathematical computer model to investigate groundwater flows toward Lake Ellesmere from the Canterbury plains aquifers. Such models give results that are only as good as the geological and aquifer properties used in the calculations.

Groundwater outflows change less with changes in lake level than do river outflows. Hence closed-basin lakes tend to exhibit much larger fluctuations in lake level in response to changes in weather and climate than do open-basin lakes. Lakes Rotoma, Rotoehu, Okataina and Rotomahana in the Rotorua Lakes district are all closed-basin lakes with groundwater inflows and outflows, while neighbouring Lakes Rotoiti, Rotorua and Tarawera all have surface outlets. An example of larger variations in lake levels of the closed-basin lakes is illustrated in the 1970-75 records for Lake Tarawera and Lake Rotoma (Figure 17.9).

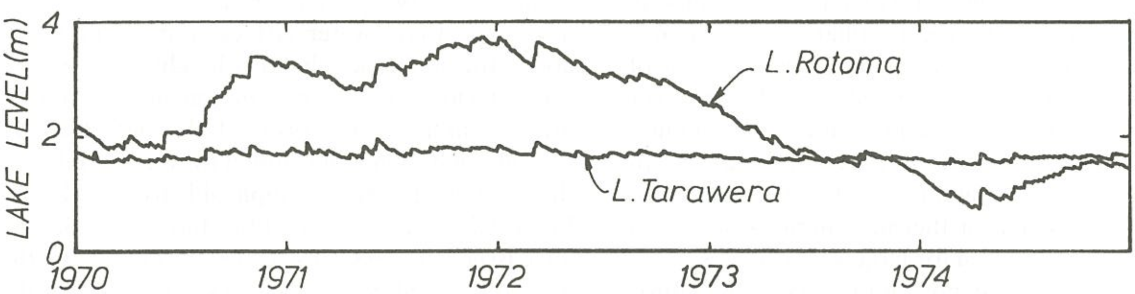

Figure 17.9 Lake surface elevations in L. Tarawera (an open basin) and L. Rotoma (a closed basin) from 1970-75; elevations are relative to a datum 277.315 m above sea level.

During this period, levels in closed-basin Lake Rotoma varied by 3.012 m, while levels in open-basin Lake Tarawera varied by only 0.477 m. The variation in Rotoma levels correlates well with rainfall. Rain (measured at Taumana near Lake Rotoma) for the six months June - November 1971 totalled 1.531 m, while that for the six months October 1973 - March 1974 was less than one-third this amount, 0.492 m.

Evaporation

Evaporation is an important term in the water balance of most lakes for time spans of weeks, months or longer. There are no published studies of New Zealand lake evaporation, although lake evaporation has been estimated from evaporation pan measurements in water balance studies by Pittams (1968) and Spigel (1989). In many arid regions of the world evaporation from lakes and reservoirs represents the loss of a valuable resource and efforts have been made to estimate evaporation rates accurately as part of water management schemes (e.g. USGS 1954). In New Zealand's humid climate, however, interest in lake evaporation has been low.

The water balance itself can be used to estimate lake evaporation. In principle this is done by treating E as the only unknown in Eq. 17.1 and measuring every other term. In practice this is often not feasible for lakes or reservoirs because of the difficulties associated in measuring the other terms in the water balance with sufficient accuracy.

Alternatives to the water balance for calculating evaporation include the thermal energy balance for the lake, mass-transfer methods, and use of evaporation pan measurements. The Penman (or Penman-Monteith) equation, which is a combination of mass-transfer and energy balance approaches, is not suitable in lakes because it assumes no change in thermal energy storage, an incorrect assumption for large water bodies.

Evaporation depends on the relative humidity, temperature, wind-speed and intensity of turbulence in the air above the water. Drier air, faster wind speeds and more intense atmospheric turbulence all increase evaporation rate. The effects of air and water temperatures are a little more subtle. Cold (thus denser) air under warmer, lighter air (caused by a cold water surface under a warm air mass) is statically stable and the stability suppresses vertical mixing. Warm water under cold air creates instability, and enhanced vertical mixing can be easily seen in the steam rising from rivers, ponds and lakes on cold, frosty mornings. All these factors can be incorporated into formulas that relate them to evaporation rate. These are called mass transfer formulas (the *"mass"* refers to water) and are usually of three types (Stull 1988): (1) *"bulk transfer"* formulas are based on the difference in humidity between the water surface and one point in the air above the water; (2) flux-gradient relations utilize a series of measurements of humidity (a humidity *"profile"* to quantify humidity gradients) at different elevations in the air above the water; and (3) eddy-correlation measurements combine measurements by fast response sensors of turbulent fluctuations in humidity and vertical wind speed. The correlation between these turbulent quantities is the vertical flux of water vapour, i.e. the evaporation rate. Only the eddy-correlation method actually measures vapour-flux directly. All other methods, including the water balance, rely on measurements of quantities other than the vapour-flux itself. Measuring eddy-correlations is both difficult and expensive and is not practical for most routine hydrologic studies.

Bulk-transfer formulas are all of the same form as Eq. 8.1 (Chapter 8). They are usually preferred for lake studies because they require measurements at only two points - one in the water as close as possible to the water surface, and one in the air above the water, ideally at a height of 10 m but often at a lower elevation (6 m, 3 m and 2 m heights are common). Some forms of the bulk transfer formulas have been developed to suit special conditions. One that may be applicable to the S. Island hydro-lakes was developed by Harbeck (1962; see also Brutsaert 1982 p.171) for reservoirs in the western United States. It takes into account the *"oasis effect"* of a water surface surrounded by arid land - the transport or *"advection"* of hot, dry air over the lake from the surrounding countryside increases lake evaporation. The smaller the lake, the more important this effect will be.

Energy balance methods utilize the fact that for every kilogram of water evaporated approximately 2.45×10^6 joules of thermal energy are required to transform the water from a liquid to a gas. This quantity, the *"latent heat of vaporization"*, relates the volume or mass of water evaporated to a thermal energy flux or heat transfer from the water to the atmosphere. Since every kilogram of water evaporated effectively carries with it an amount of thermal energy equal to the latent of heat of vaporization, the *"evaporative"* or *"latent heat transfer"* from the water to the atmosphere is

$$H_E = \rho_w \lambda E \qquad (17.3)$$

where H_E = evaporative or latent heat transfer rate per unit surface area of water (joules /(m^2 - s) = watts/m^2 in mks units), λ = latent heat of vaporization (2.45×10^6 j/kg for water at 20°C), E = evaporation (m/s) and ρ_w = density of water (approximately 10^3 kg/m^3).

Latent heat flux does not raise the temperature of the air. Heat is transported in *"latent"* form, as distinct from a *"sensible heat flux"* that arises from contact of warm water with cold air and does cause air temperature to rise. Transfer of sensible heat from the water to the overlying air occurs by molecular conduction but more importantly by turbulent convection associated with wind or an unstable atmosphere. Sensible heat will be transferred from the air to the water if the air is warmer than the water. Latent heat can be transferred from the air to the water only if the humidity of the air is higher than the water surface humidity, in which case condensation rather than evaporation takes place and latent heat is released to the water.

The Energy Balance and Some of its Implications

Both latent and sensible heat fluxes form part of the overall energy balance for a lake:

$$d\varepsilon/dt = (R_n - H_E - H_C)A_S + H_G + H_Q \qquad (17.4)$$

where ε = thermal energy (joules) stored in the lake; $d\varepsilon/dt$ = rate of change in stored thermal energy (watts); R_n, H_E, H_C are the exchanges with the atmosphere of net (incoming minus outgoing) radiation, evaporation, and convective or sensible heat, respectively, (all watts/m^2, with H_E and H_C taken as positive for upward heat transfer); A_s = lake surface area (m^2); H_G = heat transfer to the lake through the ground, including geothermal heat conduction (watts); and H_Q = net heat transfer (watts) to the lake by inflowing and outflowing streams that are at different temperatures from that of the lake. This last term also includes heating by geothermal fluids or steam, and can also incorporate heating or cooling by groundwater and rainfall if necessary - i.e., any heat that is carried, or "advected", by fluid entering or leaving the lake. Heat fluxes and storage effects associated with the energy balance are illustrated in Figure 17.10. Further information on energy balance may be found in Hutchinson (1957, pp.512-525), Anderson (1954), and Pinsak and Rogers (1981).

The energy budget can be used to compute evaporation by measuring every term in Eq. (17.4) except H_E and then solving the equation for H_E (Anderson 1954). In practice there are usually too many terms with too much uncertainty associated with their measurement, with H_E being calculated as a small difference between these larger and uncertain terms. Nevertheless, in some cases energy balance calculations for H_E can be quite effective, especially where H_E is large and the other terms can be measured carefully.

Evaporation Pans

Evaporation pan measurements do not give an accurate estimate of lake evaporation: neither the energy balance components nor the mass transfer characteristics of an evaporation pan are the same as those of a lake. Nevertheless there are many circumstances in which evaporation pan measurements are the only practical alternative for estimating lake evaporation.

These include reservoir capacity design problems for which the lake does not yet exist, and problems in which lake water temperature data and/or required meteorological data on wind speed, air temperature and humidity are unavailable. The ratio of lake evaporation to pan evaporation is typically in the range 0.6 to 0.8 for raised pans and 0.8 to 0.98 for sunken pans (Linsley *et al.* 1975). This ratio is known as the *"pan coefficient"* and must be multiplied by pan evaporation measurements to give lake evaporation. Linsley *et al.* describe methods for using meteorological data to calculate pan coefficients; these methods should be followed if at all possible when using pan measurements to estimate lake evaporation.

Lake Stratification

Roughly 90% to 95% of all incoming solar radiation penetrates the water surface. Underwater, the sunlight is used either by plants for photosynthesis or absorbed by the water itself as heat. Most of the light is absorbed in the upper part of the water column; the amount of light transmitted downward decreases steadily with depth. As a result lakes exhibit *"thermal stratification"* during summer, with a warm upper layer, relatively well mixed by the wind during the day or by cooling of the lake surface at night, overlying colder, quiet water at depth. The upper and lower layers, known as the *"epilimnion"* and *"hypolimnion"* (Figure 17.10), are separated by a transition layer in which

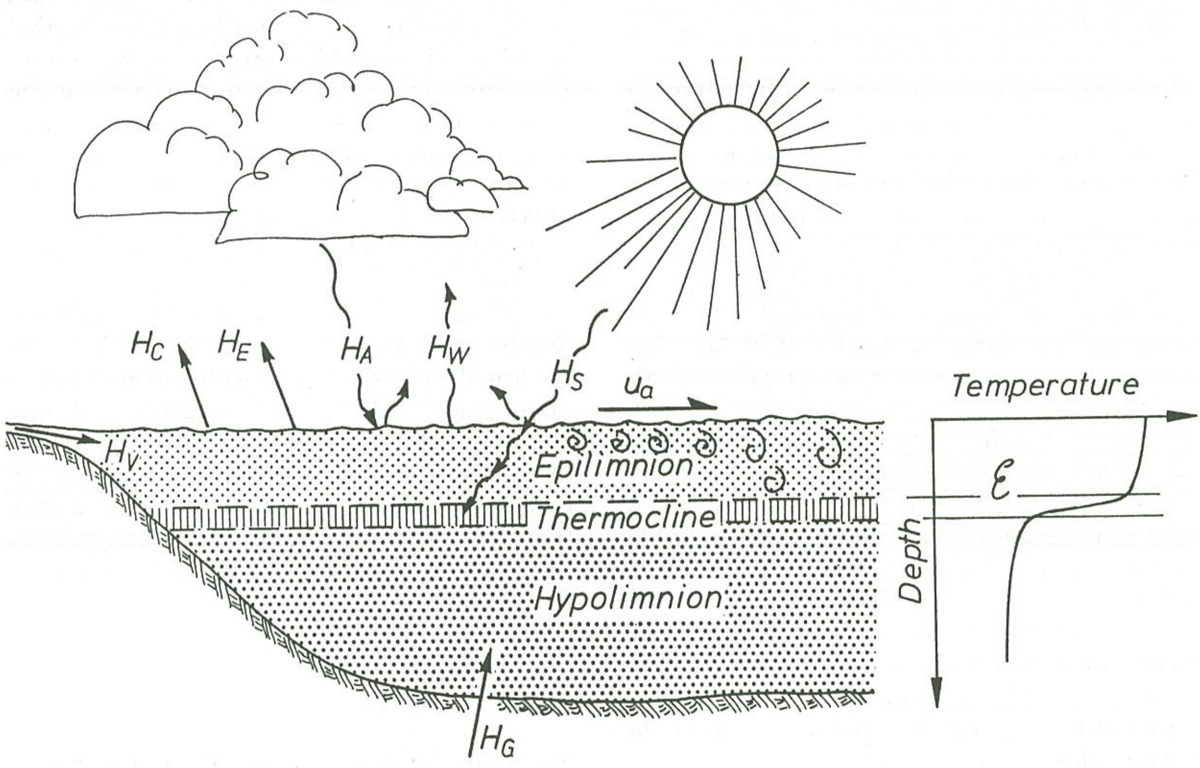

Figure 17.10 Components of the thermal energy balance. Net radiation R_n is composed of incoming minus reflected solar radiation H_S, plus incoming minus reflected longwave radiation from the atmosphere H_A, minus longwave radiation from the water surface H_W. Heat is lost by evaporative cooling H_E and convective (or sensible) heat transfer H_C. Heat can be lost or gained by advection in river flows H_V or by conduction through the bed of the lake H_G. Most solar radiation penetrates the water surface but is absorbed with depth, resulting in storage of heat $\check{S}$ in a warmer, mixed, upper layer (epilimnion), a cooler, quiet, bottom layer (hypolimnion), and a transition layer (thermocline or metalimnion).

temperature decreases rapidly with depth (*"thermocline"* or *"metalimnion"*). The relative depths of these layers depends on the transparency of the water as well as on climate, exposure of the lake to wind and lake morphometry (Davies-Colley 1988). In New Zealand, seasonal stratification is normal for lakes deeper than 10 m (Viner 1984).

Density stratification, resulting from thermal stratification, has a strong influence on mixing, circulation, water chemistry and ecology. The stratification is stable and any movements of cold, heavier water upward or warm, lighter water downward are met with resisting buoyancy forces that tend to return water to its original, undisturbed position. These restoring forces permit the occurrence of *"internal waves"*, periodic oscillations of water parcels about their state of neutral or static equilibrium. The buoyancy forces also inhibit vertical turbulent mixing to such an extent that in summer the thermocline acts as a barrier to downward transport of oxygen from the epilimnion into the hypolimnion, and to upward transport of dissolved nutrients from the hypolimnion into the epilimnion.

Internal waves can be observed as periodic changes in velocity or temperature within a stratified water column. Basin-scale standing internal waves - *"internal seiches"* - arise in much the same way as do surface seiches, when water that has been moved by wind returns to its undisturbed equilibrium position when the wind stops (Imberger 1979 pp.157-158; Wetzel 1975 pp.100-116). Temperatures measured continuously during March 1986 in Lake Rotoiti (N. Island) at depths of 13.8 m, 18.2 m, and 22.4 m (Figure 17.11a) exhibit such oscillations. During that time the undisturbed thermocline extended from approximately 16 m to 20 m (Figure 17.11b). The oscillations have a period of 21.2 h and were recorded as the thermocline moved up and down past the fixed temperature probes. Wind data were not available, but it is likely that the movements are associated with a basin-length internal seiche. The periods of internal waves are generally much longer than those of surface waves because the buoyancy-induced restoring forces are so much weaker than those due to gravity alone. Internal waves do not cause net vertical transport or mixing of dissolved substances. Unless they become unstable and break, internal waves cause only reversible displacements of fluid. Turbulent and molecular *"diffusion"*, small-scale mixing that is irreversible, are fundamentally different processes. Turbulence is strongly suppressed in the vertical by density stratification.

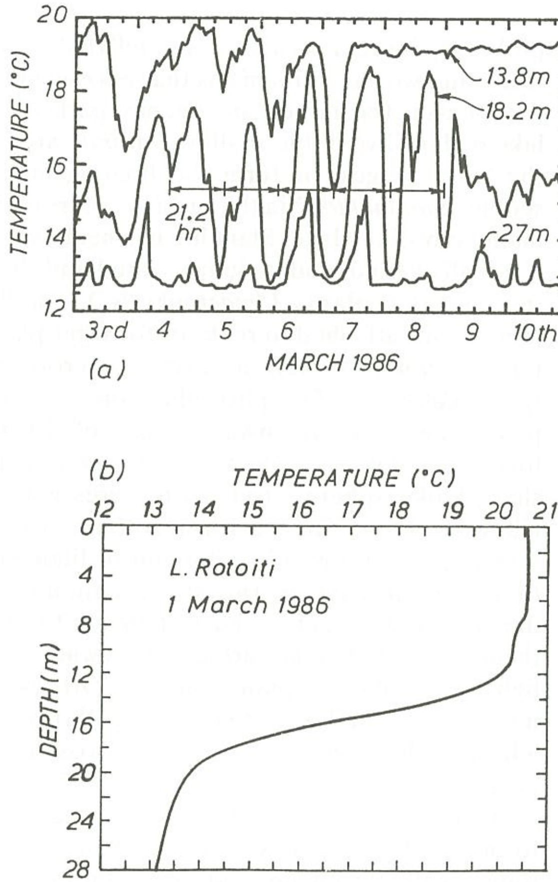

Figure 17.11 (a) Temperature data measured continuously with time in March 1986 at three depths (13.8 m, 18.2 m, 22.4 m) in L. Rotoiti (N. Island), showing periodic (21.2 h) oscillations of the thermocline. (b) The thermocline extended from 16 to 20 m and the oscillations in (a) are probably associated with a basin-length internal seiche.

Lake Ecology

Lakes provide habitat for fish, waterfowl and a vast array of smaller aquatic animals including amphibians, insects, crustaceans, jellyfish, molluscs, and worms - organisms that burrow in the sediments in the darkest and deepest parts of the lake bed, or live in the shallow *"littoral"* around the lake's edges, or form the floating microscopic *"zooplankton"* in the surface waters of the main body of the lake. Plant life in a healthy lake is equally varied, and includes rooted and floating vascular plants (*"macrophytes"*), smaller plant forms attached to rocks or to larger plants (*"periphyton"*), as well as floating microscopic *"phytoplankton"*. The phytoplankton are composed mainly of freshwater algae of diverse forms and colours. Many exist as single cells, shaped like spheres, rods, stars, crescents or ribbons. Some species form colonies or assemblages, or grow in long threads or filaments. Many are passive in the water column and depend on wind-generated turbulence to keep them suspended in the surface layers where sunlight is available for photosynthesis. Others can move using small appendages (flagella), or by adjusting their buoyancy through the loss or gain of stored carbohydrates.

In too great an abundance some algae can form unwelcome scums, cause smells and other water quality problems, and - when they die and decay - deplete a lake of its supply of dissolved oxygen, thereby literally choking other forms of aquatic life. Such overproduction of plant life occurs when there are plentiful nutrients, mainly nitrogen and phosphorous, dissolved in the water and available to fertilize plant growth. These nutrients are washed into lakes by streams and rain, although nitrogen can be obtained (or *"fixed"*) from the air by some algae and bacteria in the plankton. Once nutrients have entered a lake they can be utilized for growth by living organisms, and then recycled by processes of death, decay, and mixing. Some fertilization or enrichment (*"eutrophication"*) is natural and forms part of the life cycle of all lakes, even in undisturbed and undeveloped catchments. However, eutrophication can be greatly accelerated by human activities that increase nutrient supplies to lakes. Examples of human-induced nutrient loading include runoff from fertilized farmland, erosion from construction and road-building activities, release of sewage effluent (treated or untreated) into a lake, or seepage into a lake of groundwater that is contaminated with septic tank or landfill effluent.

Phytoplankton form the basis of the food chain in any lake, and their composition, abundance and productivity probably reflect the state of health of a lake (from a human perspective) more than any other single factor. Phytoplankton require both sunlight for photosynthesis and nutrients for growth. Light intensities sufficient to sustain net photosynthesis occur mainly in the epilimnion, thereby restricting living plants to the upper part of the water column. As summer progresses nutrients are gradually depleted in the epilimnion by phytoplankton. Because of stratification nutrients generally cannot be resupplied from the hypolimnion to match the depletion rate. Unless the shortfall is made up by runoff from the surrounding catchment, plankton production in the epilimnion will be limited. In many North Island New Zealand lakes algal populations achieve their most rapid growth in winter, when stratification is weak or nonexistent, but decline during the summer (Vincent 1983). This is quite different from the classical description of algal populations in North American and European lakes; there lakes freeze over in winter and algal growth is limited by cold temperatures and lack of sunlight, rather than by lack of nutrients and stratification. Temperatures in North Island lakes seldom drop below 8°C in winter and there is always enough solar radiation to sustain growth.

During summer dying plankton sink out of the epilimnion, thereby increasing the water clarity in the upper water layers. At the same time, the decay of sinking plankton uses oxygen that is dissolved in the hypolimnion, often depleting the oxygen at a rate exceeding downward diffusion of oxygen from the well-aerated epilimnion (Vant 1987). In summer, dissolved oxygen concentrations in the hypolimnia of eutrophic lakes often drop below levels that are necessary for survival of fish, leading to the death and eventual disappearance of some fish species from these lakes. Trout, for example, thrive only under conditions of cool temperatures (< 20°C) and high concentrations of dissolved oxygen (> 6 mg/L), and cannot survive when concentrations are persistently less than 2.5 mg/L (Rowe 1987 p.122). Their requirement for low temperatures often restricts trout to the cooler waters of the hypolimnion of most lakes in summer, but they are then susceptible to stresses of low oxygen levels.

In some lakes oxygen disappears altogether during summer from all or part of the hypolimnion. The hypolimnion is then said to be *"anoxic"*. This allows the release of nutrients into solution that had, under oxidized conditions, been locked in inaccessible form in the sediments (Wetzel 1975 pp.198, 220, 247). During winter mixing these dissolved nutrients become available to plankton for further growth. Many other important biochemical transformations depend on the levels of dissolved oxygen present in the hypolimnion (Downes 1988).

The effects of inflows, stratification and high algal productivity on the ecology of Lake Rotoiti (North Island) have been described by Vincent *et al.* (1984, 1991) Priscu *et al.* (1986) and Gibbs (1992). The lake deteriorated from the 1950's, when the hypolimnion remained well oxygenated throughout the summer, to 1970, when the hypolimnion became anoxic briefly, and finally to 1984, when the hypolimnion remained anoxic for more than three months. The dependence of dissolved oxygen on stratification is illustrated by the comparison between late winter and late summer conditions in Lake Rotoiti in 1981-82 (Figure 17.12). At the end of the winter the water column was completely mixed with respect to both temperature and oxygen. Stratification began to develop in late September and persisted through mid-May of the following year. By mid-March, however, the entire hypolimnion had become anoxic.

The high algal productivity and resulting anoxia in Lake Rotoiti puzzled New Zealand scientists for several years. The lake is relatively deep (mean depth 33 m, maximum depth 122 m) and its catchment is largely undeveloped (49.8% native forest, 42.8% pasture, 8.3% pasture and scrub - Nairn 1975), with no obvious source of nutrients to explain the increased productivity of algae in the lake in the late 1970's and early 1980's. Lakes of similar size, exposure and catchment land-use lacked similar algal productivity and summer anoxia.

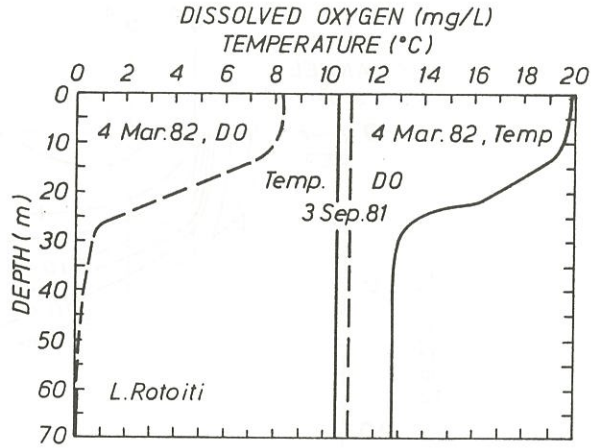

Figure 17.12 The dependence of dissolved oxygen concentrations on stratification in Lake Rotoiti (N. Island). Winter conditions of complete mixing (temperature 10.5°C, dissolved oxygen 11.0 mg/L) are illustrated by the profiles for 3 September 1981. The results of summer stratification are illustrated by the profiles for 4 March 1982; by the end of summer the entire hypolimnion was practically anoxic (Priscu *et al.* 1986; Gibbs 1992).

The only major inflow to the lake is from nearby eutrophic Lake Rotorua via the Ohau Channel (Figure 17.2 and 17.8). The channel inlet is adjacent to Lake Rotoiti's outlet in the shallow western basin of the lake, and it was thought that Lake Rotorua water affected only the western basin of Lake Rotoiti. In 1983, however, it was discovered that in winter the temperature difference between colder Ohau Channel water and warmer Lake Rotoiti water causes Ohau Channel water to form an underflow that reaches the main, deep eastern basin of Lake Rotoiti. This underflow is almost certainly the main source of enrichment for Lake Rotoiti. The movement of colder water downslope along the lake bed is

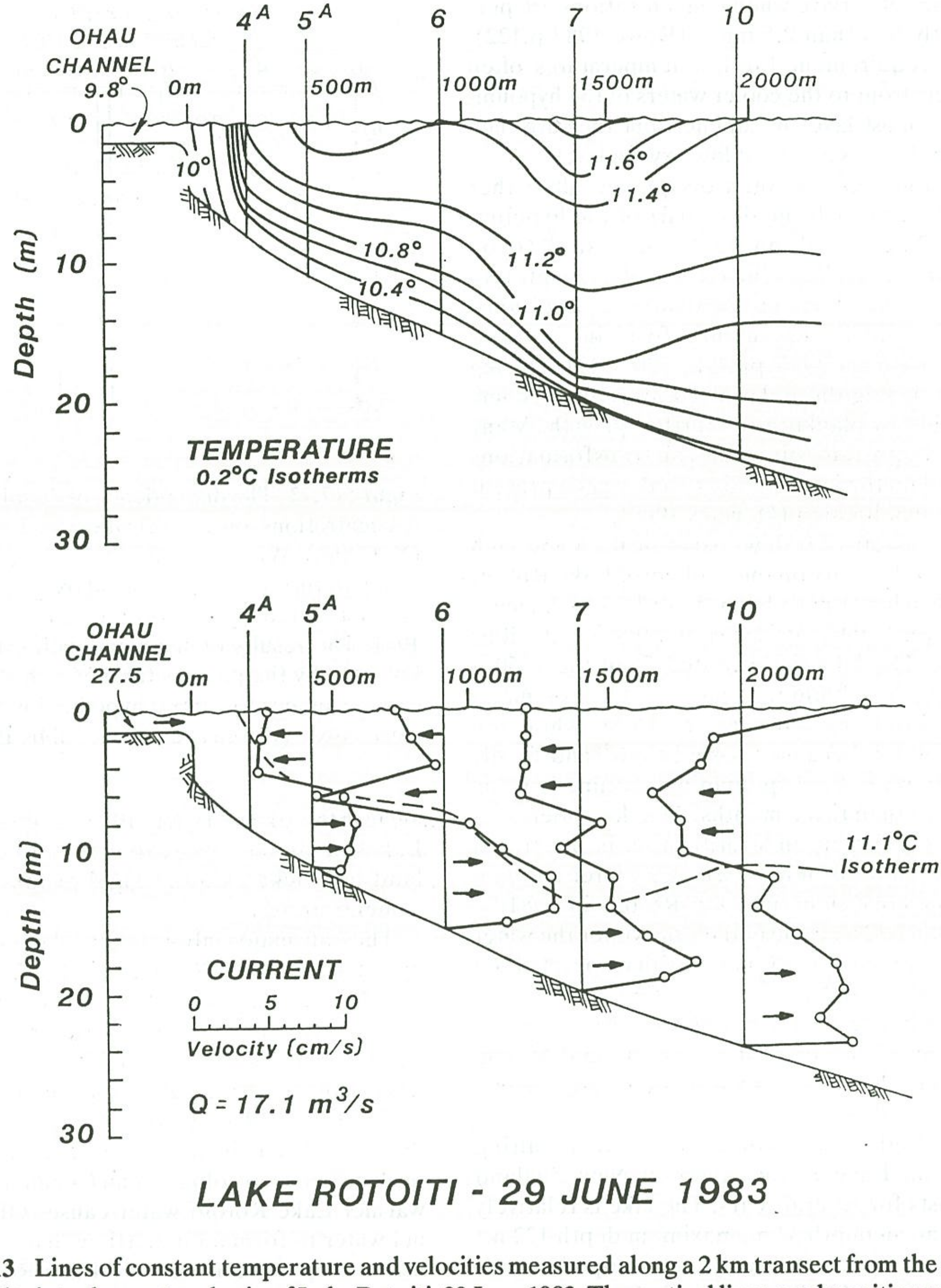

Figure 17.13 Lines of constant temperature and velocities measured along a 2 km transect from the Ohau Channel delta into the western basin of Lake Rotoiti, 29 June 1983. The vertical lines mark positions where temperature and velocity profiles were measured. The discharge from the Ohau Channel was 17.1 m^3/s and the water speed on the delta was 27.5 cm/s.

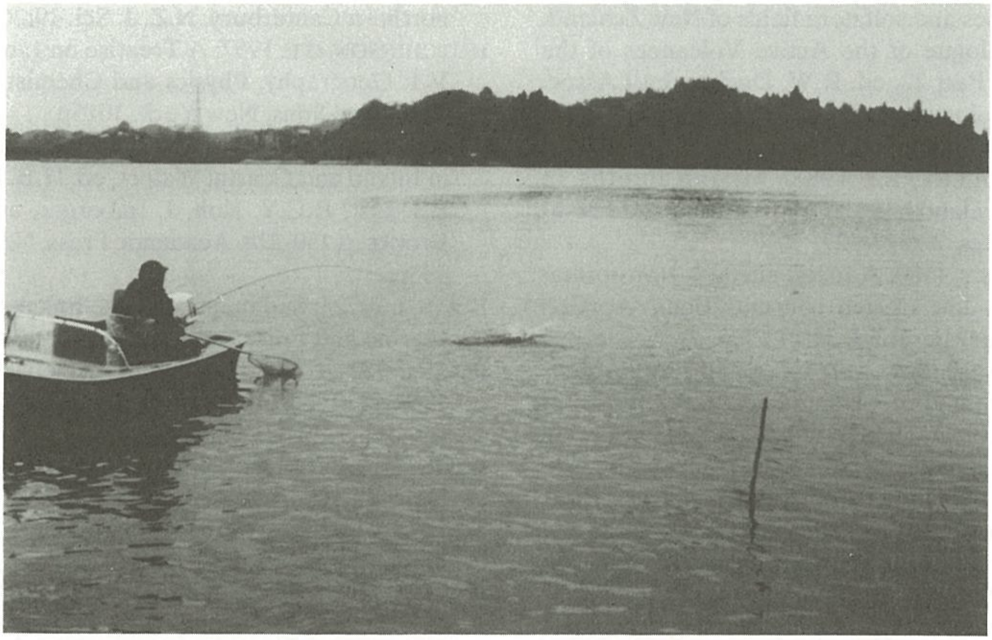

Figure 17.14 From a boat moored on the delta of the Ohau Channel, a fisherman brings in a trout; the photo was taken on the same day as the measurements in Figure 17.13. The slick (approximately 200 m from the delta) marks the area where the river and the lake currents converge and mix and the extent of the plunge region.

shown clearly by temperature contours (isotherms) and velocity profiles measured along a 2 km transect from the Ohau Channel delta into the western basin of Lake Rotoiti (Figure 17.13). The plunge region can also be detected in lake surface conditions (Figure 17.14).

A full understanding of the conditions in Lake Rotoiti was possible only as a result of studies by a group of workers of the hydrology, hydrodynamics, chemistry and biology of the lake.

Acknowledgements

We thank the National Institute of Water & Atmospheric Research Ltd, Freshwater Division, for assistance with lake-level data, and Works Consultancy - Power Engineering for assistance with reservoir morphometric data. We also thank Val Gray for help with illustrations and Pat Roberts for word-processing.

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18
Urban Hydrology

J A McConchie

Introduction

Water is the most important natural resource. It is essential for all areas of human activity. It is required for survival, drinking, food production, communication, energy generation, manufacturing, transportation, and waste removal, and is increasingly considered a recreational necessity. During the past few hundred years our need for water has increased with our "level of sophistication" until it is now the dominant factor influencing economic development and sustainability, as well as the health and well-being of society. Most decisions society makes affect our water resources. A new plastics factory, for example, requires 1000-2000 tonnes of water for each tonne of plastic produced, and most of this water is returned to the environment as polluted effluent. The irrigation necessary to grow wheat in a dry zone requires 1500-3000 tonnes of water for each tonne of wheat. In 1970 the population of the United States needed 200-500 l/head/day of water, but this was rising rapidly and could reach 1000 l/head/day (that is, one tonne of water per person each day) by the year 2000. Eighty percent of this water is used to dilute and transport sewage and industrial waste, and a similar figure could be expected in New Zealand. While the hydrological cycle is fundamentally a "natural system", it has become a technological and social system as well. For example, ten percent of the national wealth of the United States is invested in structures designed to alter the hydrological cycle in some way, e.g., collecting, diverting, storing, or distributing water (Dunne and Leopold, 1978).

While water affects human activity (Figure 18.1), human activity itself can affect both the quality and quantity of water in the urban environment. The conversion from rural to residential or commercial land uses alters the timing, volume, and risk posed by storm runoff, and this affects groundwater recharge. Urbanisation also alters sediment yields, water quality, recreational potential, and stream aesthetics and can lead to changes in channel form. Furthermore, communities require sustainable water supplies and waste disposal systems. Urbanisation is therefore intimately linked to the hydrological cycle. A sound understanding of the impacts of urbanisation is essential to both the maintenance of our quality of life and the quality of the water resource, and to reduce the costs resulting from poor management.

Changes to Patterns of Flooding

The hydrograph, which is simply a graph of streamflow through time, provides an excellent means of investigating some of the changes that result from urbanisation, particularly those relating to the volume and timing of streamflow (Figure 18.2). This is because the hydrograph reflects a wide range of processes that lead to streamflow generation e.g., the rainfall event, hydraulic

Figure 18.1 A 24-hour storm in Wellington severed communications to the Hutt Valley and caused $20M of damage from flood waters and silt deposition. (Photograph: Courtesy of The Evening Post)

parameters of the stream, and runoff characteristics of the basin.

Urbanisation causes major changes which affect runoff characteristics (Figure 18.3) and include:

1. The removal of vegetation which acts as a buffer between the impact of rainfall and the soil surface. Vegetation intercepts rain, so a considerable volume of water never reaches the ground surface to make its way into a stream. The root network and litter layer increase the infiltration of water through the soil surface, and slow down and increase the length of flow paths any precipitation must take to reach a stream. A natural vegetation cover binds the soil, absorbs the impact of high intensity rainfall, and stores and uses water, reducing and delaying the volume reaching the stream.
2. An increase in the percentage of impervious ground cover. As the soil surface is paved and sealed, the amount of water that can infiltrate the surface decreases, with a consequential increase in runoff during any storm.
3. The provision of sewers and storm-water drains. These concentrate rainfall very quickly and allow rapid runoff. In urban areas, the natural drainage network may be completely destroyed (often filled in to provide flat land for buildings etc.) and replaced with an artificial "channel" system designed to accommodate particular sized storms.
4. The drainage of swamps and bogs. During storms these areas act as "ponds" and consequently slow down runoff, reducing the height of a flood peak.
5. Modifications to slope angle and form also alter stream slope and catchment area, affecting the timing and volume of runoff.

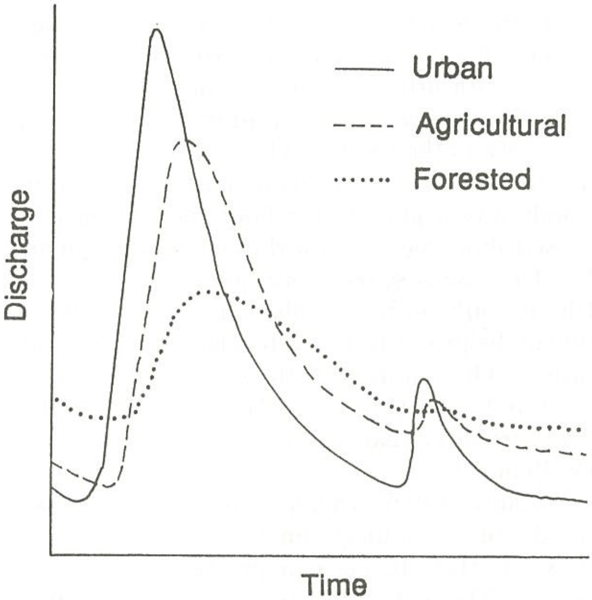

Figure 18.2 The conversion of a catchment from forest to urban land use results in more floods which rise faster and higher, but are of shorter duration.

Changes in Runoff Generation

During urbanisation floods change in response to these factors. It has been shown that the shape of a flood hydrograph is related directly to the percentage of the catchment that is impervious (Williams, 1976). As more of the basin surface becomes impermeable, the volume of storm runoff increases (i.e., there are more "floods") and the floods peak more rapidly (Figure 18.4). This latter effect is related more to the impact of artificial channels and storm-water sewers than simply the "sealing" of the surface (Hollis, 1975), although usually the two are complementary. The number of floods per year, particularly small floods, increases. Because the size of a flood is a function of the amount of rain, the storage potential of the catchment, and the amount of the catchment that is saturated, the effect of urbanisation on larger floods is usually less. During large storms most of the catchment will be saturated, and since the water cannot infiltrate the surface it will behave as if the surface had been "sealed".

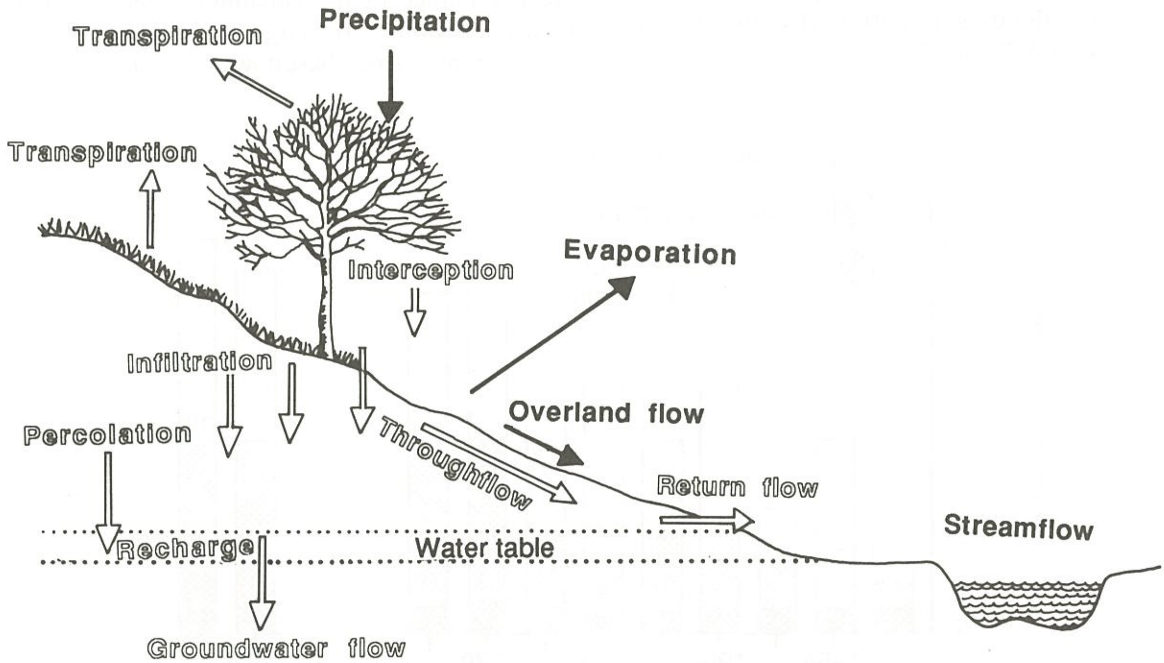

Figure 18.3 The removal of vegetation from a catchment leads to less infiltration and consequently more runoff. The soils also lose their protective buffer from the impact of rainfall and runoff and are thus more susceptible to erosion. The "open" lettering highlights those links in the hydrological cycle which are disrupted by urbanisation.

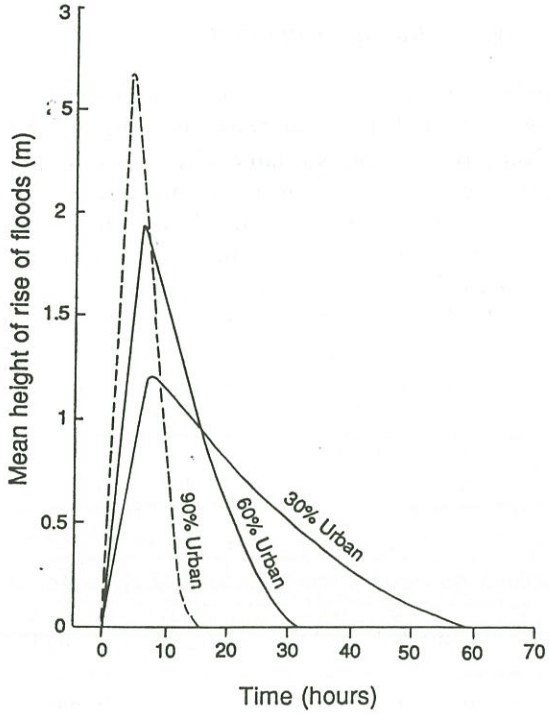

Figure 18.4 As a catchment is progressively urbanised the floods get shorter, rise higher, and strike faster (Williams, 1976).

On the North Shore in Auckland the number of small floods started to increase almost immediately with urbanisation, but the frequency of large floods increased only after 1968, by which time 53% of the basin had been developed (Figure 18.5). The increase in incidence of very large floods was negligible for both the reason discussed above and because they occur infrequently. The floods strike more quickly, rise higher (the mean height of floods increased as a function of the proportion of urban land), and run off faster. The duration of floods halved between 1963 and 1971 (Figure 18.4). As a result, the hydrograph is sharper and has a higher peak (Williams, 1976).

Because artificial channels and sewers increase the density of drainage lines and the velocity of flow, shorter storms can produce significant floods. This is because the concentration time (the time taken for water from all parts of the catchment to reach the gauging point or outlet) is reduced.

If runoff increases as a result of an increase in the percentage of the catchment that is sealed, then groundwater recharge should be reduced. Consequently, since baseflows are maintained by

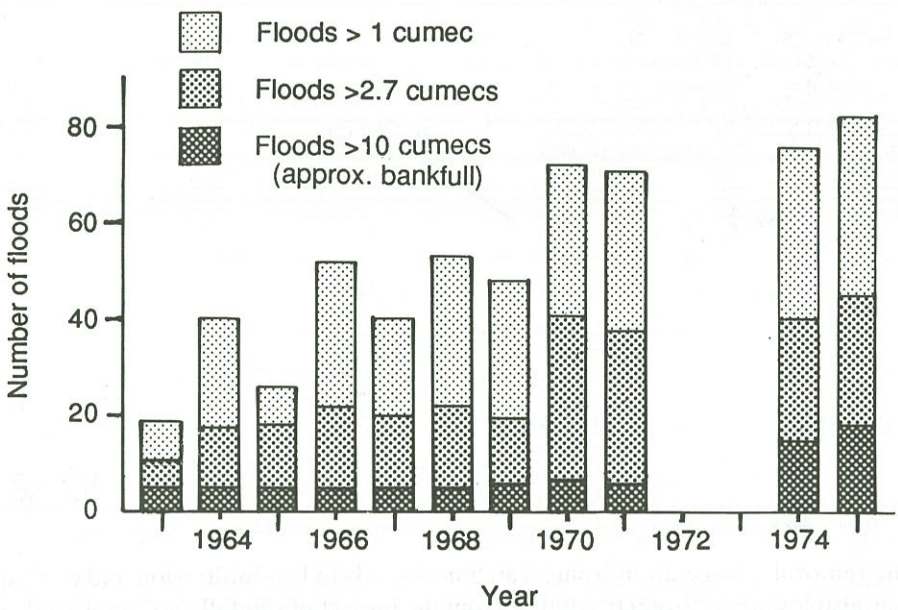

Figure 18.5 Urbanisation leads to an increase in the number of overbank (flood) flows. The most dramatic increase is in the number of smaller floods (After Williams, 1976).

discharge from the groundwater there should be a reduction in low flows. It has therefore been argued that urbanisation leads to a large increase in the variability of streamflow i.e., bigger flood peaks and lower baseflows. However, studies have shown that this is not always the case. Herald (1989) found that, although there was an increase in total discharge and a decrease in groundwater recharge, low flows also increased in both frequency and magnitude. Hollis (1977) explained the anomaly by suggesting that urban catchments respond more rapidly to lower intensity and shorter duration rainfall than a natural catchment. That is, while small rainfalls would all infiltrate and be "stored" in an undeveloped catchment, they are sufficient (when concentrated by artificial channels etc.) to generate a response in the stream, thus increasing low flows.

Changes in Channel Form

Stream channels in an unmodified catchment are in equilibrium with the amount of water they have to carry, and therefore develop a relatively stable shape and size. Since urbanisation leads to an increase in the volume of water that a stream must carry, and usually an increase in velocity, it will also lead to changes in the dimensions of the stream channel unless preventative measures are taken. Thus over time (often very rapidly) the stream will adjust its hydraulics and morphology to the new conditions, often leading to an increase in cross-sectional area. Consequently, large channels develop to accommodate the more frequent large flows, but for much of the time these channels only carry very low flows. Much land is thus "wasted" to allow for large but infrequent flood flows, and water quality is reduced.

Analysis of the Albany basin on Auckland's North Shore shows changes to stream channel morphology resulting from urban development. With no control of storm runoff a three-fold increase in stream channel cross-sectional area could be expected for a totally developed catchment. The relationship between urbanisation and channel cross-section is, however, not a simple linear function. Change occurs very slowly at first (perhaps with some threshold having to be overcome) with an increase in channel cross-sectional area of only 1.28 times by the time 50% of the catchment was developed. This increase in area reaches 1.7 times when 85% of the catchment is developed and thereafter increases very rapidly. These results do not apply to the impact of the initial earthworks on the channel, but reflect new equilibrium conditions once the system adjusts to increases in runoff and flood frequency (Herald, 1989).

Summary

Urbanisation therefore causes changes to patterns of flooding within a catchment. A study of Auckland's North Shore (Auckland Regional Authority, 1983b) summarised these:

1. The provision of complete stormwater drains and sewering of a rural basin will increase the mean annual flood (return period 2.33 years) by 1.7 times.
2. A fully urbanised catchment, completely sewered and with about 50% impervious cover, will increase the peak flow of a two-year flood more than 4 times.
3. Large floods of low frequency, such as 50 and 100-year events, are less affected by urbanisation, with their peak flows increasing by about 2.5 times.
4. The number of bank "overflows" (floods) increases - perhaps doubling where 20% of the catchment has storm sewers and is impervious.
5. Floods rise to a higher peak more quickly than under rural conditions and also run off more rapidly. This change occurs swiftly during the early stages of development. As the basin develops from rural to 20% impervious, the lag time (between rainfall and runoff) in basins of about 10 km^2 may reduce from 2 hours to 20 minutes. Thus short duration, high intensity rainfalls may produce floods, whereas under previous rural conditions a much longer duration storm would be required to produce a major flood.

6. Natural baseflow may decrease as a result of reduced groundwater infiltration and percolation after urbanisation.
7. Where the channel material is erodible, the stream channel will tend to enlarge to accommodate the larger and more frequent floods. This can undermine foundations and will produce considerable sediment, adding to the load produced by earthworks.

Sediment

Vegetation acts as a buffer between the soil and the erosive forces of the rain and runoff. The removal of vegetation, which usually accompanies urbanisation, therefore not only leads to the changes in runoff characteristics, but affects erosion rates and consequently sediment yields and transport by the streams.

Studies in the eastern United States have shown that the earthworks accompanying urbanisation can increase erosion rates from 5 to 200 times, and sediment yields in streams by comparable amounts. Data from the UK, Japan, Canada, and Mexico all suggest similar dramatic increases. Measurements made in streams and sediment traps show sediment yields ranging from 300 to 22,000 $t.km^{-2}.y^{-1}$ in basins undergoing construction. These yields are from 2 to 100 times those of the same terrain under forest or rural land uses, with the highest sediment yields being obtained from the smallest basins. This is because both the average disturbance and average intensity of rainfall are likely to be higher for smaller catchments. If these data are corrected to take account of the actual area under construction, and a sediment delivery ratio is applied, the data show erosion rates of 100 to 500 $t.ha^{-1}.y^{-1}$ from construction sites and 3 to 7 $t.ha^{-1}.y^{-1}$ from undeveloped sites (Chen, 1974).

Increases in sediment load resulting from urbanisation can have adverse effects on the environment. These include:

- □ reducing the useful life of reservoirs
- □ depositing alluvium on developed properties
- □ inhibiting recreation
- □ sealing the land surface, which further reduces infiltration and increases runoff
- □ increasing the cost and damage to flood-inundated structures
- □ reducing oxygen levels in the water
- □ inhibiting aquatic life
- □ increasing harbour and channel shoaling
- □ reducing storm sewer and drainage capacities
- □ reducing aesthetic appeal

For these reasons it is important to understand the processes which lead to sediment production and to reduce the sediment yields resulting from urbanisation.

Soil erosion by water involves the detachment of individual particles, or small aggregates, from the soil, their transport down slope, and subsequent deposition. Particles are detached by raindrop impact or runoff shear forces, and any human activities that loosen and pulverize the soil promote the process (Figure 18.6). The eroding soil particles move down slope mainly in channelised runoff, although raindrop splash on slopes can cause additional movement and increases the transport capacity of the surface runoff. The process will not start until the rainfall intensity is greater than the infiltration rate. Once runoff starts, the quantity and size of material transported is controlled by the runoff velocity and turbulence, with larger material requiring a greater velocity for movement. Because the size of particles that can be transported is a power function of the velocity, erosion and rills develop rapidly as the slope angle and runoff rates increase. An increase in sediment yields therefore often accompanies urbanisation, as a result of both the increased peak discharge during storms and the increased supply of transportable material.

Sediment Yields

The rate of urban development in New Zealand peaked in the early 1970s when in Auckland alone several hundred hectares of land were under development at one time, mainly for housing subdivisions. Investigations done as part of the Upper Waitemata Harbour study have shown that

Figure 18.6 The removal of vegetation and disturbance of the soil can lead to erosion rates of up to 200 times those under natural vegetation cover. These urban earthworks in Wellington have resulted in large rills and extensive sediment deposition in the stream channel.

sediment yield from land undergoing earthworks is about 30 times higher (25 $t.ha^{-1}.y^{-1}$) than from similar grassed land. In one North Shore (East Coast Bays) subdivision 59,000 $mg.l^{-1}$ of suspended sediment (the highest ever recorded in New Zealand at that time) was measured in one storm, even though the rainfall was not unusual. Sediment concentrations decrease and are not so dramatic as the size of the "catchment" sampled increases. This explains why the maximum recorded suspended sediment yield for the Wairau Creek in this study was only 9800 $mg.l^{-1}$ at discharge of 16 $m^3.s^{-1}$. Under natural conditions it is unlikely that sediment yields ever exceeded 500 $mg.l^{-1}$ (Williams, 1976). On the North Shore alone it was predicted that every year for the next 20 years an average of 75 ha will undergo earthworks, and that this will result in about 21,000 tonnes of sediment being deposited into Lucas Creek and subsequently the harbour each year. One stream which normally transports 150 $t.yr^{-1}$ had to cope with an extra 750 tonnes of load. Increased and more frequent floods will also lead to erosion of the stream banks, and this will supply an additional 40,000 to 100,000 tonnes of material per kilometre along small to medium streams (Auckland Regional Authority, 1983a). The deposition of sediment and consequential problems downstream are often ignored. However, in one year, 0.3 m of sediment resulting from urbanisation was deposited over the Wairau estuary (Strachan, 1977).

Such dramatic changes are, however, relatively short lived (Figure 18.7). For example, in the mid 1970s approximately ten percent of the Wairau Valley on the North Shore was under development, and annual sediment yields were about 10 times those of neighbouring rural catchments. By 1982, with urban development substantially complete, and with the main channels concrete-lined, sediment yields had reduced to about the same rate as that of the rural catchment (Auckland Regional

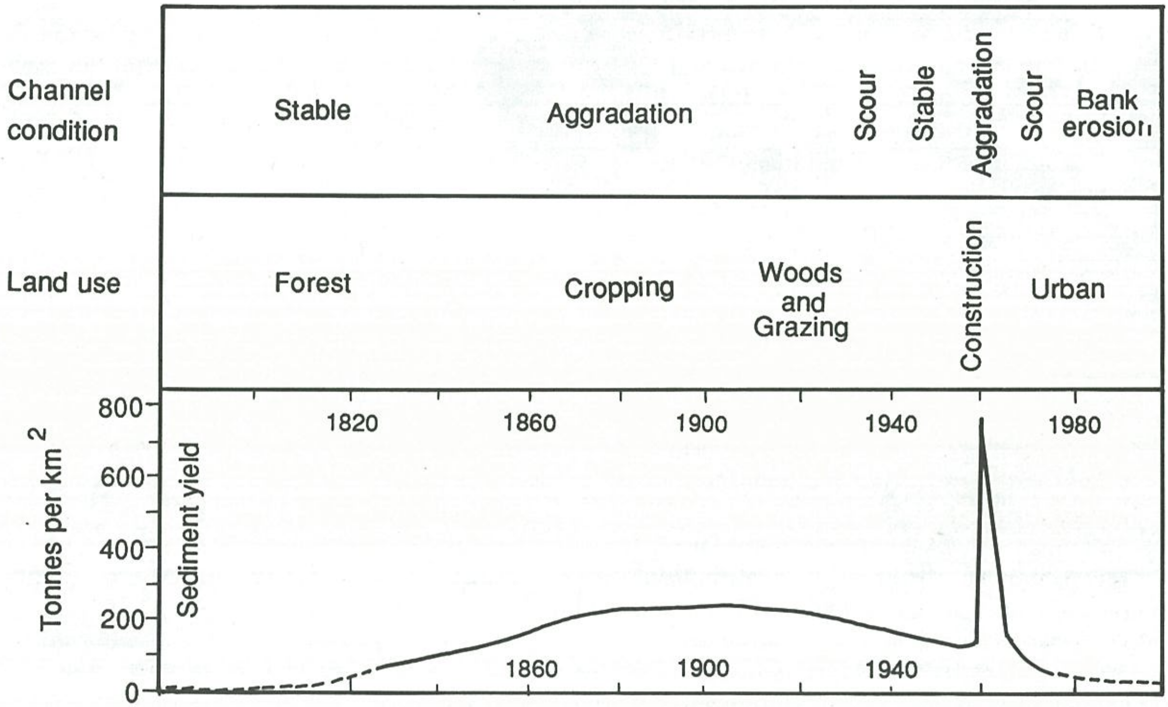

Figure 18.7 Sediment yields and stream channels respond to land use changes. The impact of the construction phase of urbanisation is, however, usually short-lived (Wolman, 1975).

Authority, 1983a). Therefore, prior to 1950 the erosion rate in the Wairau Valley was approximately 3000 $t.yr^{-1}$ and this occurred during occasional intense storms. With removal of trees and scrub this rate doubled. The highest rates recorded were from construction sites, which yielded 11 times the amount of sediment of non-urban areas even though only 21% of the area was actually under construction. Small, highly disturbed construction sites could yield more than 60 times the amount of sediment as the same area under rural land use. Once construction stops, sediment yields fall. New suburbs, for example those less than five years old, have yields only 5 times their rural equivalents. Sediment yields continue to fall and after 30 years can even be lower than the initial rural rates (Strachan, 1977).

Sediment yields resulting from urban development were also investigated in the Pauatahanui Basin, 30 km north of Wellington. The major results of this study (Curry, 1981), were that:

1. Daily sediment inputs to the Pauatahanui Inlet during the two year study period ranged from a minimum of 0.01 $t.d^{-1}.km^{-2}$ to a maximum of 85 $t.d^{-1}.km^{-2}$, with a mean of 0.35 $t.d^{-1}.km^{-2}$.
2. Annual sediment yields for the two year study period ranged from approximately 75 $t.km^{-2}$ for a low altitude pastoral catchment to approximately 160 $t.km^{-2}$ for a partially urbanised catchment. Earthworks in the Browns Stream catchment produced approximately 1200 $t.km^{-2}$, more than 15 times the yield of its pastoral counterpart.
3. Suspended sediment concentrations were generally within the range of 5 to 500 $mg.l^{-1}$; however concentrations in Browns Stream, that was undergoing earthworks, ranged from 25 to 1500 $mg.l^{-1}$.

4. The maximum recorded mean daily sediment discharge of the Browns Stream was approximately 1640 $t.d^{-1}.km^{-2}$, about 35 times greater than its pastoral counterpart and 22% of the total sediment input to the Inlet on that day (Browns catchment area is 1.2% of the Inlet catchment area). For a 1 in 5-year recurrence interval flood, Browns Stream produced a maximum instantaneous sediment discharge of approximately 6700 $t.d^{-1}.km^{-2}$, about 75 times that of its pastoral counterpart.
5. The long term mean annual sediment yield was 37 $t.km^{-2}$.
6. The average annual flood (approximately 1 in 1-year recurrence interval) is estimated to contribute 12% of the long-term mean annual water yield and 50% of the long-term mean annual sediment yield.
7. The mean annual flood (1 in 2.33-year recurrence interval) is estimated to contribute 90% of the long-term mean annual sediment yield, while the 1 in 5, 10, 20, 50, and 100-year floods are estimated to produce approximately 150%, 200%, 300%, 400%, and 750% of the long-term mean annual sediment yield.
8. All particles of suspended matter were silt-sized or smaller (<0.06 mm) and, with the exception of Browns Stream, an average of 50% were clay sized (<0.002 mm). Suspended sediment from the Browns Stream catchment was finer, with more than 75% in the clay range. When discharged into Browns Bay with its low energy currents, the suspended particles flocculated and settled before tidal flushing could remove them from the Bay.

These contemporary rates of erosion and sedimentation can be put into historical context through examination of sediment cores from Waitemata Harbour and Pauatahanui Inlet. From the dates obtained at various intervals, together with the thickness of material between, it is possible to calculate the average rate of deposition. At the top of the Pauatahanui core, over a period of 2250 years (3610-1360 B.P.), about 5.5 m of sediment was deposited, giving a mean rate of about 2.4 mm per year. However, further down the column, at a depth of about 10 m, 1 m of sediment was deposited in 80 years (7970-7890 B.P.), a mean rate of about 12.5 mm per year. This higher sedimentation rate suggests that a large amount of debris must have been available from the catchment during this period (Healy, 1980). The impact of human occupation is obvious in a core taken from Lucas Creek in the Waitemata Harbour. European settlement coincided with a dramatic four-fold increase in sedimentation rates to approximately 3 mm/yr (Figure 18.8). This rapid increase may have resulted from the widespread destruction of vegetation (Auckland Regional Authority, 1983c). Variation in erosion and sedimentation rates can therefore reflect natural changes in the supply of sediment and the erosive potential of both rainfall and runoff. Human modifications of the landscape, however, can lead to dramatic, but often short-lived, increases in these rates.

Lucus Core

Depth m	Vegetation Change	Cause of Change	Average Sedimentation
0.00		Surface of mud	
	Introduced pollen types appear eg pine, willow plantain, and grass pollen increases	European settlement	3 mm/yr
0.4		100 yr BP	
	Bracken becomes the dominant type	Polynesian settlement and forest clearance	0.8 mm/yr
0.8		530 yr BP	
0.9			
	Catchment covered with undisturbed kauri, rimu rata, beech, and totara forest		
1.1 Core limit			
10.0 Basement not reached by probe			

Figure 18.8 The stratigraphy of sediment cores taken from the Upper Waitemata Harbour show a dramatic increase in sedimentation with the arrival of Europeans and the removal of forest vegetation (Auckland Regional Authority, 1983c).

Floods

Although it has been argued that urbanisation increases the size and frequency of floods, this does not mean that floods are either a new, or simply an urban, phenomenon. Successive flooding of the Hutt River in the 1840s, for example, led to the first settlers in Petone relocating to Wellington, and Blenheim was named "Beaver Town" because of its frequent floods. In fact, during the late 19th century flooding reached such proportions that it was dubbed "the New Zealand Death". The increase in population densities, often on flood plains or close to stream channels, however, have certainly increased the risk posed by these natural events.

Between 1920 and 1953 New Zealand experienced 820 damaging floods; so severe were the frequency and magnitude of flooding during the 1930s that the Soil Conservation and Rivers Control Act (1941) was passed with the aim of reducing both soil erosion and flooding. Losses, however, have continued to increase despite the legislation and massive investment in prevention measures. In the 30 years following 1953 there were 115 large damaging floods. Since 1970 there have been 15 major regional floods affecting many urban areas. An estimate of the direct property losses from these events approaches $1 billion ($90 million per year) and some 6,000 homes and 1,500 businesses have been directly affected. Curiously, of 17 urban areas flooded since 1970, 80% had some form of flood control (Ericksen, 1986).

Nearly 100 communities in New Zealand are currently prone to flooding. Of 136 communities with more than 1000 people in 1956, 64% had experienced one or more floods since 1920. From the early 1950s to 1984 over two thirds of these places (containing 67% of New Zealand's population) were revisited by floods, some several times, and many in spite of being protected. Nearly 70% of New Zealand towns and cities with populations in excess of 20,000 have river flood problems. While entire urban centres are not likely to be inundated, of the 18 communities examined in detail in 1981, the average built-up area lying within the historical flood zone was 20%. While this is staggering, many of these damaging floods were not very rare events, being in the 20 to 80-year return period range, and this has severe implications for the future (Ericksen, 1986).

Millions of dollars have been invested in flood protection schemes which have focused on the physical process and engineering structures, such as stopbanks, to reduce flood losses. While this has reduced the number and frequency of floods, Ericksen (1986) argues that it has actually increased the flood hazard and the actual disasters in urban communities. He suggests that flood hazard (the probability of occurrence of a potentially damaging natural phenomenon) is as much a human creation as a consequence of the flood itself. Indeed, flood hazard is shown to be a function of both the flood event and human use of the floodplain - including the adjustments people make to reduce flood losses. If the area flooded had no use, and contained no people, then it would not be a problem!

Overflows onto a flood plain (flooding) are common natural events which over the years, through deposition of sediment carried at high flows, have led to the formation of the flood plain. Human occupation of the flood plain does not stop this geomorphic process. It should be expected therefore that some parts of low-lying flood plains and valley floors will be inundated every 2-3 years and the rest, unless uplift or changes in the hydrologic characteristics of the basin occur, could be inundated by large floods with return periods of 50 or 100 years. That is, in any one year these areas have a 2 or 1 percent chance respectively of being flooded.

Wide differences exist between the perceptions of floods by "experts" and by the "lay public". Experts expect floods in areas where they have already occurred, and the longer the time span the larger the possible event. They therefore describe floods in terms of their return period. This confuses many people who think that if a 100-year flood has occurred one year it will not do so again for 100 years. In fact, a 100-year flood has a one percent chance of being equalled or exceeded in any year. People also find it difficult to imagine future events much larger than those previously experienced and often seem to take comfort in the fact that the most recent event is the biggest

(Ericksen, 1986)! Most people also believe that stop-banks will stop flooding, when in fact although engineering works are designed to contain a certain size flood they will be overtopped by greater flows. Engineering works also represent a compromise between cost and level of protection, and it is usually economically impossible to provide protection against the "maximum possible" flood.

A small rural town, Paeroa (population 3,702 in 1981) suffered property losses of around $8.5 million during one flood in April 1981. At the height of the emergency nearly 50% of the town was inundated (Figure 18.9), over one-third of the population (1,300 people) were evacuated, and 544 homes were flooded (219 inside, some to a depth of 2 metres).

Although the rainfall that generated the flooding was rare, the flow of the Ohinemuri River which overtopped and breached the stopbanks was not. It had a return period of about 70 years - that is, it has one chance in seven of occurring every 10 years. This was not the first flood in Paeroa. Flooding in 1907 and 1910 led to the adoption of the flood control scheme which failed in the flooding of 1936 and 1954. Since then Paeroa has been threatened several times, most recently in 1976. Although the Government had approved the new Waihou Valley flood control scheme in 1971, Paeroa's portion of it was incomplete in the April 1981 floods because of local opposition to financing the scheme.

In spite of this flood history, and expert predictions about future floods, the Paeroa Borough Council continued its policy of permitting new housing and commerce in low-lying portions of the town. Much of this development was made possible by loans from the Housing Corporation - a Central Government agency. By continuing to invade and occupy the low-lying floodplain, the flood

Figure 18.9 Paeroa during the height of the 1981 floods. (Photograph: Courtesy George Caddie, HCB)

hazard in Paeroa grew throughout the 1960s and 70s and thereby the potential for flood losses (Ericksen, 1986).

Although insurance against flooding is relatively cheap, almost 40% of buildings flooded in the Thames-Paeroa region in April 1981 were not insured and many more were under-insured. Ironically, considering the cost of repairs (probably about $7.2M), almost all re-development after this flood took place in the same floodable locations. That is, people saw the flood as "a chance in a million" and were prepared to take the risk. While improvements to the stopbanking will reduce the risk of flooding, properties and businesses still have a one-in-ten chance of being flooded by the year 2000 (Ericksen, 1986).

Mitigation

There are generally three ways that we attempt to reduce flood losses. We attempt to:

1. Modify the flood: usually through engineering works such as building dams or stopbanks, or by channelling the river.
2. Modify flood losses: through emergency actions, restoration funds, and insurance.
3. Modify the human use of the floodplain: through land-use planning and zoning, and building style changes such as elevating the foundations.

Most of the measures in New Zealand have, until recently, focused on the first two approaches but losses have continued to rise. Engineering works are designed to fail in some event, and as our length of occupation of New Zealand has increased, so has the likelihood that a flood will exceed the design capacity. Also the data available when many of these schemes were designed was poor, which made it difficult to predict accurately the size of possible floods. Option two does little to change the losses as it only attempts to minimise the impact and stress. It could be argued, in fact, that the measures adopted so far in New Zealand with respect to floods have actually increased the risk and losses, and that the situation many communities now find themselves in is partly a result of past "protection" strategies.

Option three offers the greatest possibility of reducing the flood risk in the longer term. It is based on the principles that flooding is a natural process, that it will occur, that there is no "perfect" protection, and therefore we must modify our behaviour to accommodate floods when they occur. Underpinning this is a philosophical change from believing that we can control and dominate the environment to one where we accept that we are just one part of a dynamic system. Because of historical attitudes and existing land use, however, this approach is often not possible without considerable disruption, and therefore engineering structures provide a short-term solution to give time to make the necessary human adjustments.

The Engineering Approach

The traditional approach for mitigating the effects of flooding has been to build stopbanks and other containment structures to stop the water from going where it is not wanted. If these are the only methods available then the best solution is to detain stormwater in small volumes as near to its source as possible (Figure 18.10). This water can then be released at a controlled rate. This reduces the flood peak, although it will lengthen the time over which higher flows are experienced.

Porirua Stream drains about 54 km^2 and traverses 12 km from headwaters to the sea. Floods have been recorded regularly since 1846. In recent decades urbanisation of the catchment has increased runoff, and at the same time development near the stream has led to an increase in the cost of damage caused by floods. Damage during the December 1976 storm dramatically illustrated the inadequacies of the flood channel during large events.

A 100-year storm would directly affect about 66 ha of the Porirua basin, flooding properties to a depth of up to 1.5 m. The total capital value of property that would be flooded is approximately $153 million. The flood would affect 260 businesses (which employ 1640 people) and flooding of some electrical substations would cut power to

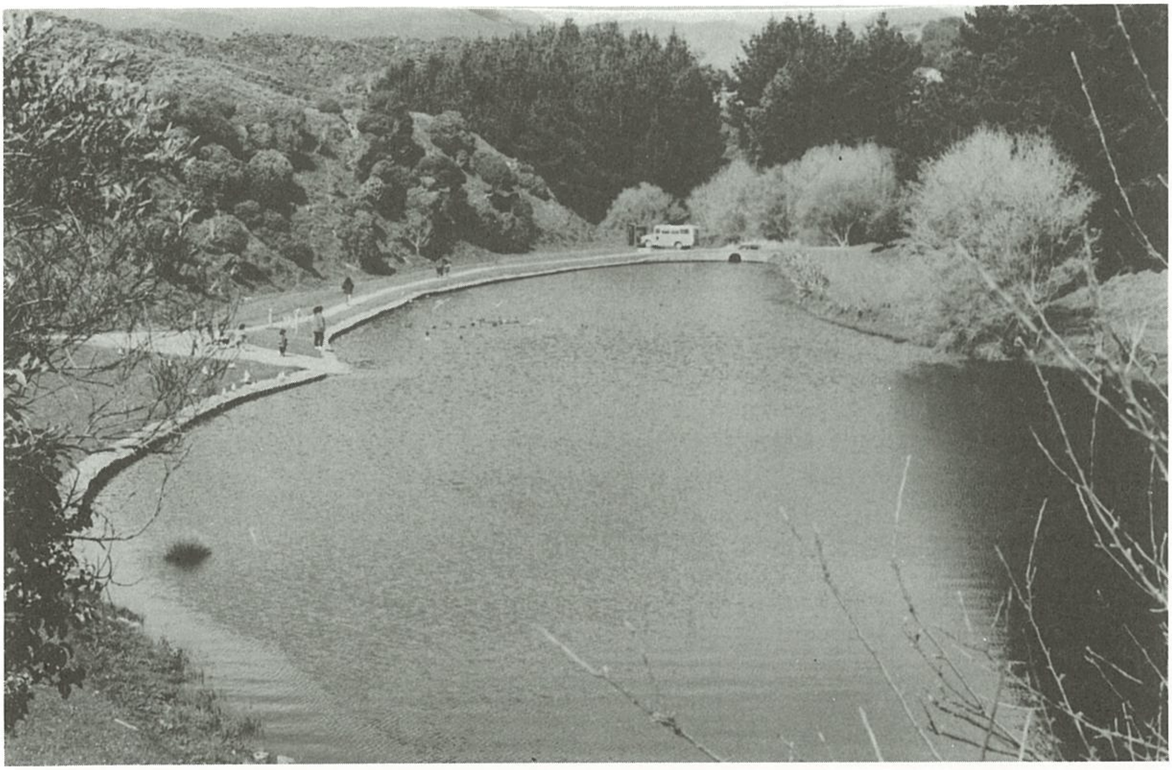

Figure 18.10 The most effective way of reducing the impact of floods in urban streams is to construct detention dams near the head of the catchment. This dam in Porirua is used as a recreational reserve under normal conditions.

most of the Porirua-Tawa urban area. Flood waters would also affect sewerage pumping stations, various main access routes, and the NIMT railway (Figure 18.11). Four hundred residential properties would be flooded: in some cases only the gardens would be affected, but a majority would have floodwater under or in the house, some up to 2 m in depth. As new houses frequently have concrete floor slabs and low ground-to-floor levels, they would incur greater damage from flooding (Wellington Regional Council, 1989).

Options available to reduce the risk were restricted because of the existing land use along the valley floor, and because much of the land close to the stream banks was privately owned. As a result engineering approaches were the only ones available and even these were constrained. The scheme adopted is based on two approaches and will be implemented gradually over the next 20-30 years to spread the capital cost ($14-16M). One is to "store" a portion of the flood runoff in large detention dams in the head of the valley and the other is to improve the channel hydraulics to carry the flood waters.

Channel improvements include widening and grading the stream banks to an optimal slope, straightening the stream course where necessary, and removing obstacles to the flow of flood water. One section (about 20% of the stream length) will be left in a "natural" state primarily to provide the friction necessary to slow the flow of floodwaters downstream, and because any flooding of this area would not cause significant damage. Where sufficient space is available, a trapezoidal earth channel (with 27° banks) sown in grass is proposed which will blend with the existing natural features

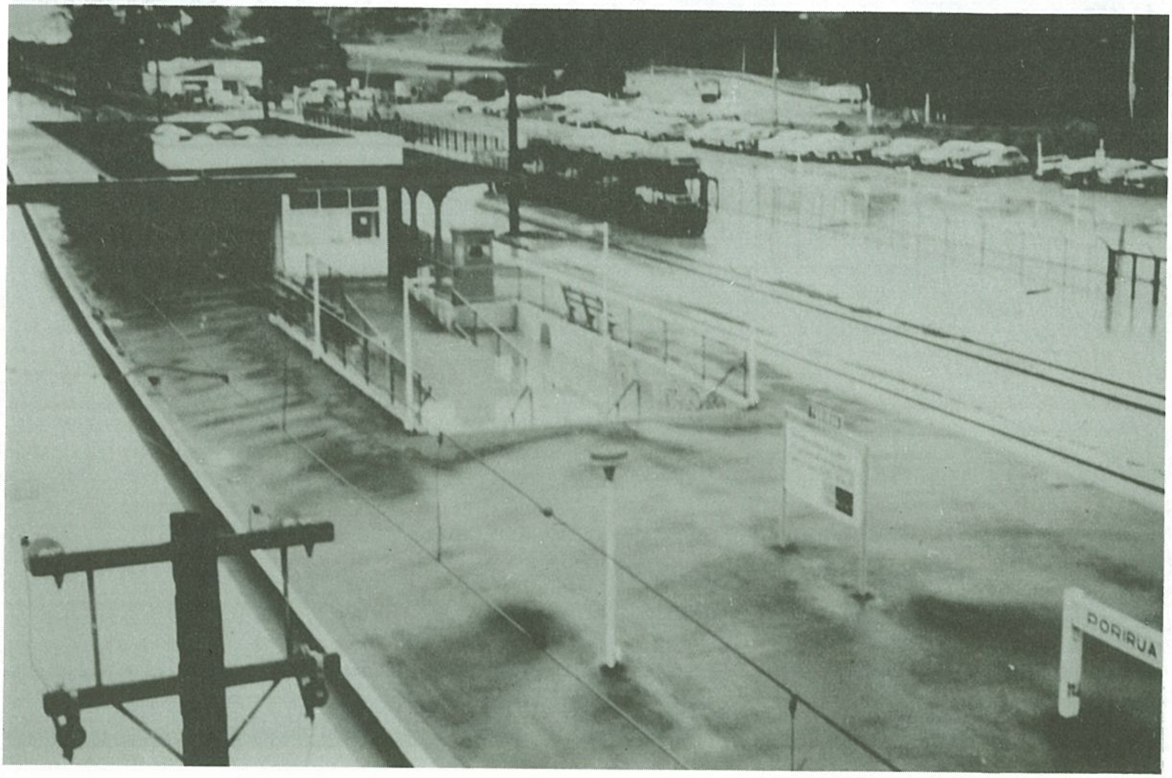

Figure 18.11: The 1976 floods in Porirua cut most of the main transport links including the NIMT railway. The Porirua station was isolated and the pedestrian subway filled with water. (Photograph: Courtesy Porirua City Council)

and is the cheapest option. Where space is not available, because of residential development or private ownership, a concrete-lined channel will be used, allowing steeper banks (35°) and faster flows.

One of the detention dams will be constructed in Belmont Gully, an upper subcatchment, and this will retard approximately 10% of the flow. At the mouth of Belmont Gully the stream flows through a culvert under State Highway 1 and the highway embankment will be used as the dam. This embankment is over 30 metres high and approximately 200 wide at the base. Since the embankment was not designed, or constructed as a water retaining structure, it has been exhaustively tested. The large culvert under the motorway has been modified to restrict the flow during floods and this will allow runoff to pond upstream of the embankment. Careful analysis of the probable maximum rainfall has shown that, provided the culvert does not block, the embankment will never be overtopped. Even with complete blockage of the culvert, there is sufficient storage available to accommodate the estimated 10,000-year flood.

Most engineering structures are designed for either the size of a flood of a particular return period or for the intensity of rainfall of a duration equal to the concentration time of the catchment. While models can accurately predict the water component of storm runoff, it is much harder to take account of sediment yields and debris. This is often ignored, or dealt with simply by arrestor screens to stop debris from entering the storm runoff system. Many of the larger storms which generate floods also generate a large volume of debris which either blocks the storm water control structures or overloads them. For example, in Stokes Valley during an intense rainstorm in December 1976, 78 landslides, mostly in the heads of stream channels, displaced about 33,500 m^3 of debris which choked the channel downstream and caused widespread flooding (McConchie, 1980). It is therefore important that this component of storm runoff is considered in any measure designed to minimise the impact of floods.

Quality

Urbanisation affects not only the quantity of water but also its quality. Changing land use, and the development of urban areas, can lead to changes in both the physical and chemical parameters of water.

Physical Properties

The most important physical characteristics affecting water quality are sediment concentration and temperature. Sediment becomes a pollutant when it exceeds natural levels and has a detrimental effect on the biologic and aesthetic values of water. While every stream carries a natural amount of sediment depending on its catchment characteristics, sediment concentration often increases markedly as a result of urban land uses. The stream may no longer appear "clean" and clear but be discoloured for long periods of time. The effect of sediment on the flora and fauna of a stream is complex. In some streams the fine silt and clay will clog the interstices of a gravel river bed, reducing the availability of "nesting" sites for many types of fauna and destroying the habitats of others. Increases in sediment concentration also lead to a rise in turbidity, affecting the depth to which light will penetrate. Elevated sediment concentrations, through reducing light penetration, may affect the efficiency, or even survival, of species that require photosynthesis. Reducing turbidity, however, does not always improve water quality. In streams with a high natural sediment load, reducing turbidity may lead to increases in photosynthesis and consequently algal blooms.

The temperature of water in streams can also be altered by urbanisation. The removal of overhanging vegetation increases the variability of temperature and this can lead to stress in many organisms that have low temperature tolerances. Also the widening of channels, either naturally or for flood control, lower baseflows, and the input of waste water can all lead to increases in water temperature. The effects of raising water temperature are many. An increase in temperature lowers the viscosity of water and therefore causes suspended matter to settle faster. In some situations this may be an advantage, but it can also mean that sediment is deposited in the channel rather than being carried further down stream. Increases in temperature also decrease the solubility of oxygen while increasing rates of oxidation, which places an increased demand on the limited oxygen remaining. Since high levels of oxygen are essential for a healthy stream, reduced levels can promote the growth of less desirable organisms and place stress on other species.

Chemical Properties

The natural chemical composition of water is a function of the rock type of the catchment supplying groundwater and of the relative ratio of groundwater to rainwater. Trace metals such as iron, zinc, lead, and copper are released in small quantities by rock weathering, but at higher concentrations in streams they can be toxic or can disrupt aquatic ecosystems. For example, industrial effluent in an English river with a copper concentration of 1-2 $mg.l^{-1}$ exterminated all animal life for 16 km downstream. The effect could be observed 30 km below the outfall, even after dilution to 0.1 $mg.l^{-1}$ of copper (Dunne and Leopold, 1978). Concentrations of trace metals are often increased through urbanisation, particularly where industrial areas lie close to watercourses and lack adequate measures to prevent effluent entering streams. Even if the effluent is collected in a sewer, metals and other contaminants can kill the bacteria used in waste-water treatment plants. Landfills also tend to concentrate many of these metals, particularly iron, lead, and zinc, and often other harmful contaminants. If the landfill is not adequately contained, these contaminants can leach into the groundwater or local streams, polluting the water. Since it may take many years for this pollution to be detected, it may become very expensive or impossible to clean up. Even when all the leachate from a landfill is collected, the disposal problem remains. Sometimes it is discharged via an industrial sewer, shifting the problem down stream and at other times it is returned to the landfill. In both cases, the concentrated leachate poses an additional risk to the environment.

Problems caused by poor land zoning are illustrated in data from the Waiwhetu Stream in Petone near Wellington. Roger (1978 a & b) described the stream as an "open drain for wastes" and divided the blame for pollution between the industries contributing the waste and the lax application of legislation by the Hutt River Board.

Evidence presented to the Town and Country Planning Appeal Board in 1976 stated that only two organisms survived in the lower reaches of the stream, one a hardy crab and the other an estuarine worm. The stream was described as "dead" with high levels of toxic metals such as lead, nickel, and zinc in the muds. The pH of discharges into the stream were as high as 12.4 at NZIG's plant, affecting the pH of the stream both above and below the discharge site as a result of the tidal nature of the river.

In 1978 an industrial sewer was commissioned to accept all discharges. This immediately reduced visual pollution associated with the Feltex carpet factory (discharging dye house residues) and the Ford Motor Company (discharging water from the wet paint sanding processes) although the waste was now simply discharged directly to the sea at another location. A year later, Davis (1979) reported that, although flora and fauna were virtually non-existent in the tidal reaches of the stream, schools of small fish were moving upstream on the incoming tide. The levels of metals in the sediments had not been reduced as a result of discharges ceasing 12 months earlier.

Fundamental to a healthy aquatic ecosystem is a high level of dissolved oxygen. The concentration of dissolved oxygen can be reduced, and its variability increased, through urbanisation. Increases in temperature can adversely affect the solubility and availability of oxygen. Also urban channels are often smoother, with fewer rapids to aid re-oxygenation. More importantly, however, the amount of organic matter entering the stream which must be broken down (oxidised) increases.

Every stream has a certain capacity to break down, or assimilate, waste matter. If this capacity is exceeded then stress will be placed on the ecosystem, leading to either subtle changes or a complete failure of the system. The oxygen required to break down any biodegradable pollutant will depend on the amount of the pollutant and its chemistry, or more particularly the mass of oxygen required to oxidise a unit mass of the substance to a stable state. This "pollutional strength" is measured by an index called the Biochemical Oxygen Demand (BOD). The BOD of a waste is therefore the amount of oxygen consumed by living organisms (mainly bacteria) feeding on the organic matter in the waste. For example, it takes 0.076 kg of oxygen to assimilate the waste produced by one person each day (Dunne and Leopold, 1978). Because of the complexity of the processes which utilise dissolved oxygen, the BOD is usually defined as the amount of oxygen used by the organisms in a sample kept at 20°C in the laboratory over five days. For domestic sewage, and many industrial wastes, about 70-80% of the total BOD is exerted within 5 days. Some representative BOD_5 values are given in Table 18.1.

Effluent Source	BOD_5 at 20°C (mg/l)
Distilling	10,000 - 30,000
Pulp and paper	20 - 20,000
Wool scouring	200 - 10,000
Canning industry	400 - 4,000
Meat packing plants	600 - 2,000
Dairy processing	200 - 2,000
Breweries	500 - 1,250
Untreated domestic sewage	100 - 400
Urban storm runoff	>10

Table 18.1 Representative values of BOD_5 for effluent produced by various manufacturing processes (Dunne and Leopold, 1978).

Urbanisation often increases the inputs of various salts, including nitrogen and phosphorus to streams. These two in particular are required for vegetation growth and elevated levels often lead to algal blooms and the excessive growth of aquatic weeds. These are responsible for the weed problems being experienced in many lakes (e.g., Rotorua) in New Zealand.

Although all these chemical contaminants exist naturally in rivers, urbanisation often leads to higher levels and a greater degree of variability. For example, a 12-hour storm in the Wairau basin on the North Shore produced: 500 tonnes of particulate matter; 4 tonnes of organic waste (BOD); 800 kg of total inorganic nitrogen; and 200 kg of total phosphorus (Figure 18.12). Predictions of

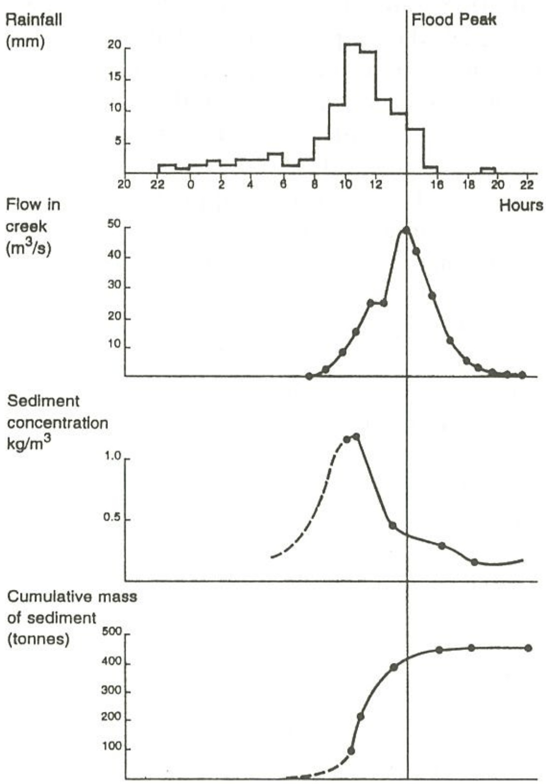

Figure 18.12 During the storm of 15 March 1980 Wairau Creek on Auckland's North Shore carried 500 t of sediment, 4 t of organic waste, 800 kg of inorganic nitrogen, and 200 kg of total phosphorus (Auckland Regional Authority, 1983d).

increased nutrient, organic, and sediment loads that could be flowing into the Upper Waitemata Harbour annually by the end of the century as a result of urban landuse include a 16% increase in total phosphorus, a 9% increase in total inorganic nitrogen, a 13% increase in BOD, and a 33% increase in sediment (Auckland Regional Authority, 1983d). Since these values are increases over the natural levels they will have significant, but difficult to predict, effects on the ecosystem.

The legacy of past land-use decisions certainly compounds the risk of polluting urban water courses. In November 1990 an accidental spillage of alkyd resin occurred at a plant near the Porirua Stream as the material was being transferred from processing tanks to a transport trailer. Alkyd resin spilt onto the concrete yard and flowed directly into the stream via the stormwater system.

When the alkyd resin reached the stream it emulsified, forming a white, very sticky material. This floated downstream adhering to rocks, vegetation, and the banks of the stream. A hydrocarbon film leached from the resin and floated in pockets of calmer water. The emulsified resin coated the feathers of ducks, and as they attempted to clean their feathers, the resin was transferred to their beaks, making it difficult, or impossible, to feed and breathe.

This problem arose largely because the factory is sited on the banks of the stream and next to a residential area. Although current planning restrictions would prevent this, the site was developed during the 1950s when there were few planning constraints. Furthermore, the site lacked facilities to prevent spillages from reaching the stream, with no means of containing stormwater on the site and no provisions for spillages in the yard to be separated from stormwater flow. As a result of this accident a strategy, involving building impermeable barriers around the storage area and interceptor drains to stop spills from reaching the stormwater system, was developed. A further spillage in 1992 has led to the firm having to spend $500,000 to upgrade its plant.

Although spillages affect urban streams, their major impact may be on the estuary and harbour where the streams discharge, and this is often ignored. A study of the sediments in Porirua Harbour attempted to identify the extent of any industrial pollution. Although not highly industrialised, Porirua City includes an automotive assembly factory, hosiery and wallpaper manufacture, together with small plants producing products ranging from foodstuffs to sheet metal products. Stormwater from this industrial area drains to the harbour in the south western corner. Results showed no evidence of contamination from cobalt, nickel, iron, and manganese. The copper concentration, however, was higher than background levels with two samples from the vicinity of Porirua City being classed on the Igeo scale (Müller, 1979) as moderately contaminated. Lead and zinc levels in this area were also higher than baseline values (Glasby *et al*, 1990).

Even in this area of relatively little industry, heavy metal pollution has still occurred in the vicinity of stormwater outlets into Porirua Harbour.

Water Supply

Although fresh water was usually a requirement when siting new towns, in many areas demand now often outstrips supply. Supplying the needs of Los Angeles, for example, requires bringing water from 320 km to the north and 400 km to the east. While the level of water use in New Zealand is significantly less, each household still consumes 600 l/day simply for domestic purposes. The history of water supply in Wellington is typical of urban areas in New Zealand, and is one of continual growth in demand and exploitation of water resources at an increasing distance from the City.

History of Supply

The Wellington region has abundant water resources, but they are located in difficult terrain a long way from where they are most needed. The early settlers got water from wells, springs, streams, and tanks and barrels collecting rainwater. When Wellington became New Zealand's capital in 1865, Parliament and the surrounding buildings were supplied with water from a spring at the foot of Tinakori Hill. This was in fact the first water-pipe laid. Such an *ad hoc* and "open" system obviously led to health problems. In 1871 Dr James Hector wrote to the City Council that *"... no water collected within the crowded parts of the city, either from wells or house tops, is safe or proper for human consumption"*. During the 1870s a water storage and distribution system, based on the Lower Karori reservoir, was built to supply a population of 7,000 to 8,000. The region's first publicly available piped water was "turned on" in 1874. This reservoir is still in use today and is the only significant storage for the City south of Upper Hutt. Although it holds 114 million litres of water, and was supposed to be "the ultimate answer to all the City's water supply problems for all time", it now represents less than two days average usage.

By 1878 water shortages were already being experienced (the supply was turned off at night) and the system required supplementing. A water storage reservoir was built in the Wainuiomata Valley (30 km east of Wellington) and a tunnel constructed under the Wainuiomata Hill in the 1880s. By 1911 even this supply was being fully utilised. Storage was therefore doubled at both the Wainuiomata and Karori reservoirs. By 1925 additional water was having to be drawn from the Upper Orongorongo River, adjacent to the Wainuiomata catchment. By the 1950s water shortages were again imminent and the Hutt River supply was harnessed at Kaitoke in 1957. This water was chlorinated and piped 56 km to the Lower Karori dam, supplying reservoirs in the Hutt Valley, Porirua, and the northern suburbs on route. Other areas had developed local water reticulation and many small streams were used for water supply by local residents. These supplies were eventually replaced following completion of the Kaitoke scheme.

In the Hutt Valley, an ample supply of groundwater (a large part of it artesian) had been discovered soon after European settlement. Wells were sunk in the 1880s, and within 50 years several hundred existed. Lower Hutt and Petone both relied on this source of water, and in 1935 Wellington began to tap the aquifer. This extensive and uncontrolled use put such a strain on the resource that in 1959 the Hutt Valley Underground Water Authority was set up to promote conservation and pollution control. This resource was developed up to the present time, and the Waterloo Pumping Station now supplies water from artesian sources to Lower Hutt and Eastbourne (Morrison and McDougall, 1986).

Not only Wellington, but most other urban centres in New Zealand, lack water storage capacity. The trend has been to develop "run of the river" type water supply schemes to provide a regular supply of naturally clean water and as demand increases, to seek "new" sources of water. This has led to water supply problems. When floods occur and the rivers carry a lot of suspended load, water extraction must stop, as the consumer would object to the colour and taste, and the plant required to pump and transport the water would become clogged and suffer excessive wear. Bacteria can also reach very high numbers in turbid water, and the "protective shelter" provided by the organic material or particles prevents chlorination from being effective. In Wellington this problem is lessened to some degree by having various supply catchments which respond to different storm conditions. In southerly storms the catchments to the east tend to flood, while in northerly conditions it is the Hutt catchment. Because of the limited storage, supply problems can still be experienced if a storm lasts for more than a few days.

Wellington's supply is also threatened during periods of low flow, such as in summer when demand can be higher than normal because of irrigation. To overcome the problems of low flow and excessive silt during floods, and to fully treat water from the Hutt scheme, two storage lakes (with a capacity of 2930 million litres) and a treatment plant have been constructed at Te Marua just north of Upper Hutt (Figure 18.13).

Traditionally, because of the high quality of water from natural catchments, water supply has

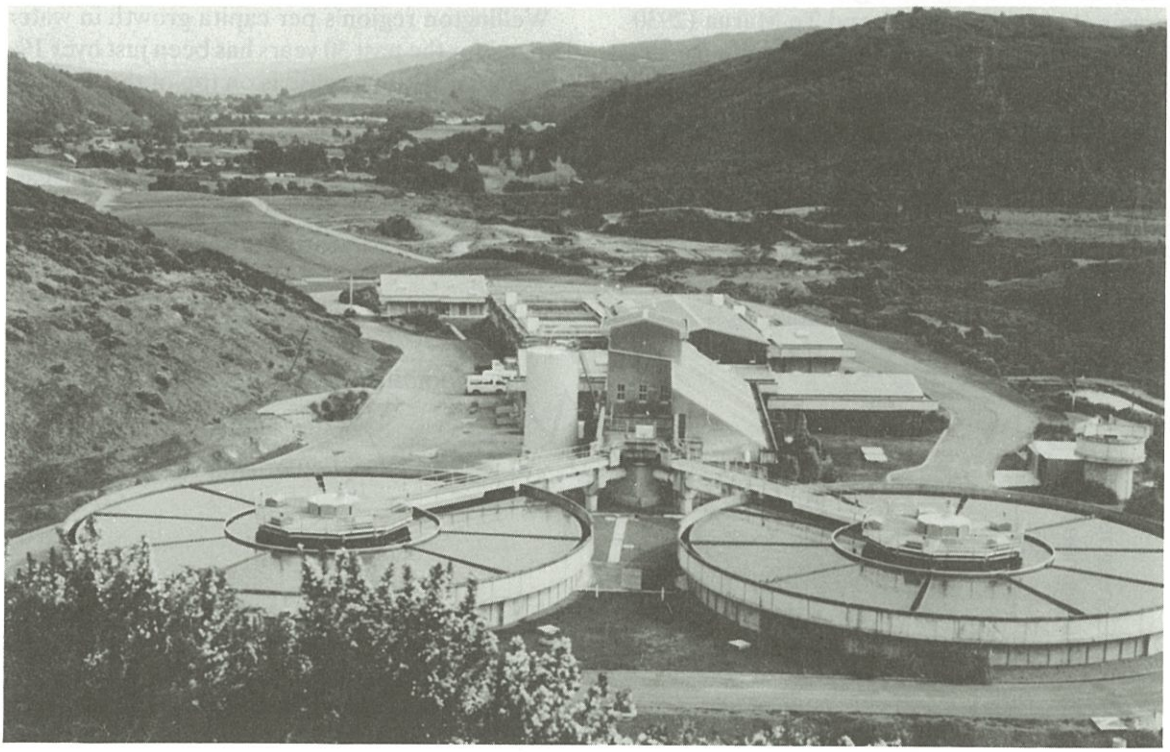

Figure 18.13 The Te Marua water treatment plant and storage lakes north of Upper Hutt provide high quality water by killing bacteria and reducing turbidity. This scheme provides the only major water storage for the Wellington region.

involved little treatment. The chemical balance and turbidity of water supplies was controlled, but sterilisation to kill bacteria was not introduced until after 1957. The modern treatment process used at Te Marua involves adjusting the pH of the raw river water so that when poly-aluminium-chloride (PAC) together with a polyelectrolyte is added, small "cotton wool" like tufts (called flocs) form. These attract any suspended matter and bacteria. The flocs, which have combined with the unwanted matter, settle to the bottom and are removed. The clarified water is then filtered through layers of anthracite coal and sand. Following filtration the water is chlorinated for disinfection, fluoridated, and corrected for final pH using caustic soda.

Although this system has supplied reasonably cheap, reliable, and excellent quality water there has always been a threat to supply because of the limited storage capacity. The only real storage capacity in the Wellington region is in the Karori reservoir (114 million litres) and Te Marua (2930 million litres). The numerous distribution reservoirs are sufficient only to meet the daily requirement of the area they serve. The bulk storage is at Te Marua, approximately 40 km to the north, and all supply pipes have to cross the active Wellington fault several times before reaching the city.

Demand

The historical pattern of Wellington's use of water has been one of continual increase in both total and per capita consumption (Figure 18.14). Growth in the per capita consumption for an average day in the week with maximum demand has been extrapolated to predict future demand. Planning at this level is essential since the supply is "run of the river" with limited storage. Since water is provided on a regional basis it is possible to smooth local supply anomalies.

Most models for predicting usage start from two main elements; estimates of future population and per capita usage. These models must be refined as the area under consideration gets smaller or more heterogeneous. The demand for water is then broken down into a number of different sectors, the factors controlling demand for each are analysed and the results are summed to produce estimates of demand.

Wellington region's per capita growth in water demand for the past 30 years has been just over 1% per annum. This increase is on top of the demand caused by an increase in population and probably results from a variety of factors e.g., increasing number of water-using appliances, water for gardening and reserves etc. A recent decline in per capita usage has been attributed to factors such as

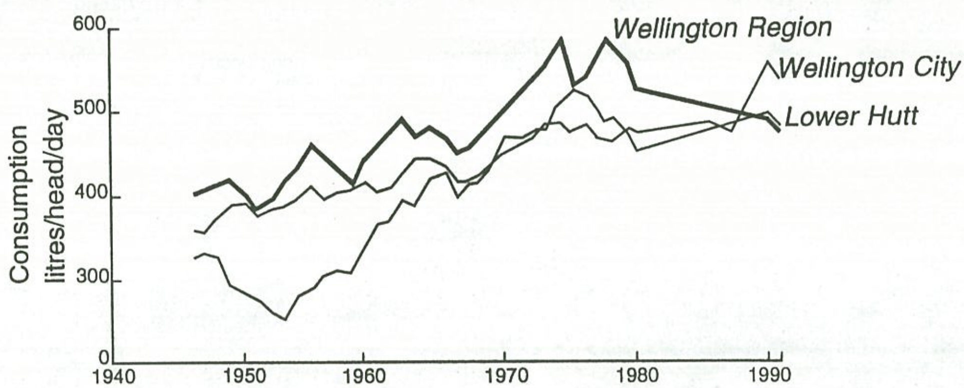

Figure 18.14 Pattern of water use in the Wellington region from 1946 to 1992 (Extended from Wellington Regional Water Board, 1980).

declining economic activity, unseasonal summer weather, and saturation of the market in the sale of appliances. The current distribution of water use is summarised in Table 18.2.

Average Use of Water	
Community	
Dwellings	39%
Industry and commerce	35%
Other	10%
Losses	16%
Domestic	
Toilet	31%
Outdoor (garden, car)	20%
Kitchen, bathroom, laundry	49%

Table 18.2 Distribution of water use, both community and domestic, for the Wellington Region. (After Blakemore, 1990).

Of the 230 l/head/day that is used for domestic purposes, the majority (80%) is used to dilute and transport sewage (Table 18.3).

Use	Litres/Head/Day
Bathroom (one 5-minute shower)	60
Toilet (five 11-litre flushes)	55
Laundry (one washing machine load may vary from 70-240 litres per cycle)	45
Kitchen (one sink half full is 7 litres. A dishwasher may use from 24-45 litres/cycle)	20
Outdoors (summer hosing for 30 minutes per day)	50

Table 18.3 Approximately eighty percent of the domestic consumption of water is used either for cleaning or to dilute and transport sewage (After Blakemore, 1990).

The amount of water used can vary with factors such as the climate, national attitudes, past experience and availability of water, so the percentage of water used in various activities differs both regionally and nationally (Table 18.4).

An improvement in living standards also leads to an increase in domestic consumption. As living standards rise, households purchase more appliances that use water e.g., dishwashers, waste disposal units, swimming pools, air conditioning, etc. or new models of appliances that use more water. For example, by 1976 45.1% of households in the Wellington region had automatic washing machines, which use more water than earlier types, and as the sales of these machines increased so did the per capita consumption (Wellington Regional Water Board, 1980).

The age structure of the population also affects water usage. Studies in Australia showed that dwellings with 2 persons used 30% more water per capita than dwellings with 4 persons, and this has been confirmed in Lower Hutt (Blakemore, 1990).

Conservation

While the demand for water is increasing, its availability is not, and new supply schemes are very expensive and logistically difficult to construct. The more efficient use of current water supplies is therefore important if restrictions are to be avoided. The two most obvious areas where savings can be made are in reducing the amount of leakage, currently about 16% of supply in Wellington, and in bathroom usage. While it is difficult to reduce leakage in the short-term, changes to water use and appliances in the bathroom can lead to immediate savings.

A water use study in Perth, Australia found that showering accounted for 78% of bathroom use, and it is likely that similar figures would apply in New Zealand. Shower consumption, however, varies with driving pressure, shower rose design, and duration. The first two factors alone can vary flow rates from 5 to 16 litres /minute (deluxe multihead models can use up to 64 l/min). Mains pressure systems can also more than double the flow rate of a shower unless flow control fittings are installed. The fitting of low-flow shower roses, or flow control valves on taps, can lead to savings of 33% of shower use or up to 25% of bathroom use (between 15 and 21 l/h/d). Baths, however,

	Australia (Melbourne)	United Kingdom	USA	Lower Hutt (NZ)
Basic (drinking, cooking, washing up)	15	26.5	14	
Personal washing, bathing, showers	38	37	25	59
Laundry	16	12.5	4	
Toilet	31	33	41	35
Sundries	(in 2 above)	3	4	6
TOTAL	100	100	100	100

Table 18.4 The amount of water used for various purposes is a function of a range of factors including climate, national attitudes, and availability of water. (After Wellington Regional Water Board, 1980).

probably use more water, depending on the length of the shower and the frequency and depth of the bath. A full bath, for example, may use 200 litres of water, although the Perth study found that average bath depth was only 116 mm (58 litres).

Toilets are the other high use appliance, using approximately 31% of domestic consumption. Most toilet cisterns hold 11 litres - that is every flush uses 11 litres of water. The trend therefore has been towards dual flush cisterns with either 5.5/11 or 4.5/9 litre capacities. Savings from the installation of dual flush cisterns can be as high as 25-33% of toilet use, or 14-18 l/h/d (Blakemore, 1990).

A considerable amount of the water used in the urban environment is used to dilute and transport waste. Disposing of this sewage is a problem for all urban communities.

Waste Water

As far back as Roman times sewerage pipe systems were used to transport waste to nearby streams and rivers, which could assimilate the small amounts of human waste with little apparent detrimental effect. Population growth, together with industrial and commercial expansion and development, mean that urban areas now produce an excessive amount of waste. Even an area with relatively little industry such as Porirua produces 370 litres of waste per person that requires disposal each day. Communities, realising the dangers and adverse effects of this waste on the environment, are installing more sophisticated sewerage treatment systems.

Water makes up over 99% of sewage because it is the transport media used for disposal. The rest is composed of small amounts of organic and inorganic matter which can be either suspended or dissolved in the water. A typical breakdown of "urban" sewage is shown in Figure 18.15.

Even though water makes up the bulk of sewage, the organic matter would accumulate if it were discharged directly into a water body. Bacteria which feed on sewage would multiply rapidly and soon use all the dissolved oxygen present. This would lead to anaerobic conditions and all normal life would disappear. The object of sewage treatment therefore is to lower the suspended solid concentration and oxygen demand by reducing its organic content before discharge. While the degree of treatment used to be determined largely by the assimilative capacity of the receiving water, social and cultural pressures now often demand significantly higher levels of treatment. While in 1950 in New Zealand there were only five treatment plants, this number has now risen to 217 and all communities with populations over 5000 have reticulated sewerage systems.

Besides the problems caused by organic

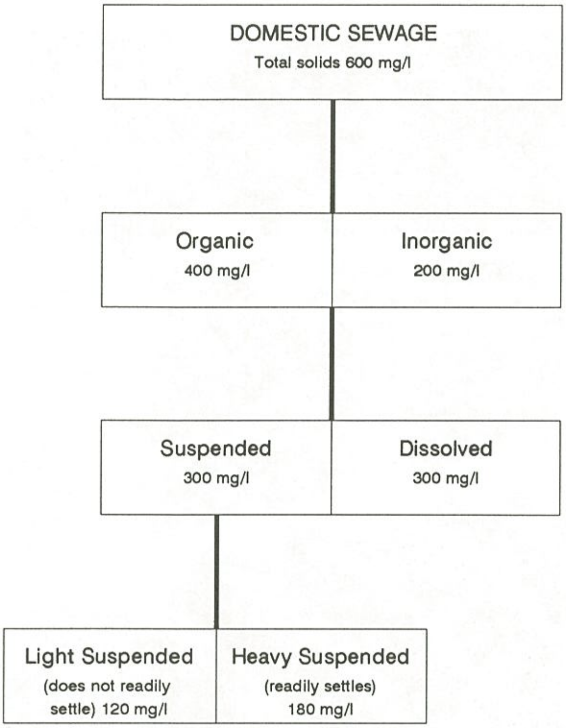

Figure 18.15 Solids content of typical urban domestic sewage (Davis, 1984).

material in sewage, inorganic salts can also have an adverse effect on the environment. For example, treated municipal sewage adds about 35 kg of inorganic salts per year per person to the receiving water (Dunne and Leopold, 1978). These salts, particularly those containing nitrogen and phosphorus, may stimulate plant growth which can clog water-ways and reduce the dissolved oxygen content even further.

Treatment

There are a range of processes available to reduce the pollutional strength of sewage. All treatment processes usually start with some form of *preliminary treatment*. In the past this consisted of mincing the sewage into small particles that could be broken down more quickly by bacteria. This technique has now been replaced by milliscreening, in which sewage is passed through a very fine sieve which removes much of the suspended solid material. The screenings are usually disposed of in a landfill, although composting options are being investigated. For many years this was the only treatment provided to sewage, and in some areas it still is, but increasingly this represents only the first stage (Table 18.5).

	Suspended solids	**BOD$_5$**	**Bacteria**
Raw waste	300 mg/l	300 mg/l	10×10^6 /100 ml
TREATMENT	Percentage Reduction		
Primary sedimentation	40-70%	15-30%	50%
High rate filters	70-90%	80-90%	95%
Activated sludge	80-95%	85-98%	98%

Table 18.5 Various sewage treatment processes are designed to reduce the suspended solid concentration, BOD, and bacteria of effluent to levels that minimise its impact on the environment (After Davis, 1984).

Primary treatment produces the "lowest" grade of treated sewage and usually consists simply of some storage facility that allows the organic solids to settle out of suspension. Because the solids settle by gravity, primary treatment removes only the larger particles, and the quality of the effluent is largely a function of the length of time the sewage remains in the tanks. The sludge which settles out can then be treated or disposed of separately, and the remaining effluent is discharged.

In *secondary treatment* aerobic biological activity is used to remove most of the remaining suspended organic matter and much of the organic matter contained in solution. In trickling filters, the effluent is sprayed over rocks where it flows in thin films over various organisms living on the rock (Figure 18.16). These organisms, which include bacteria, fungi, and protozoa (and insects which feed on them) break down organic substances in

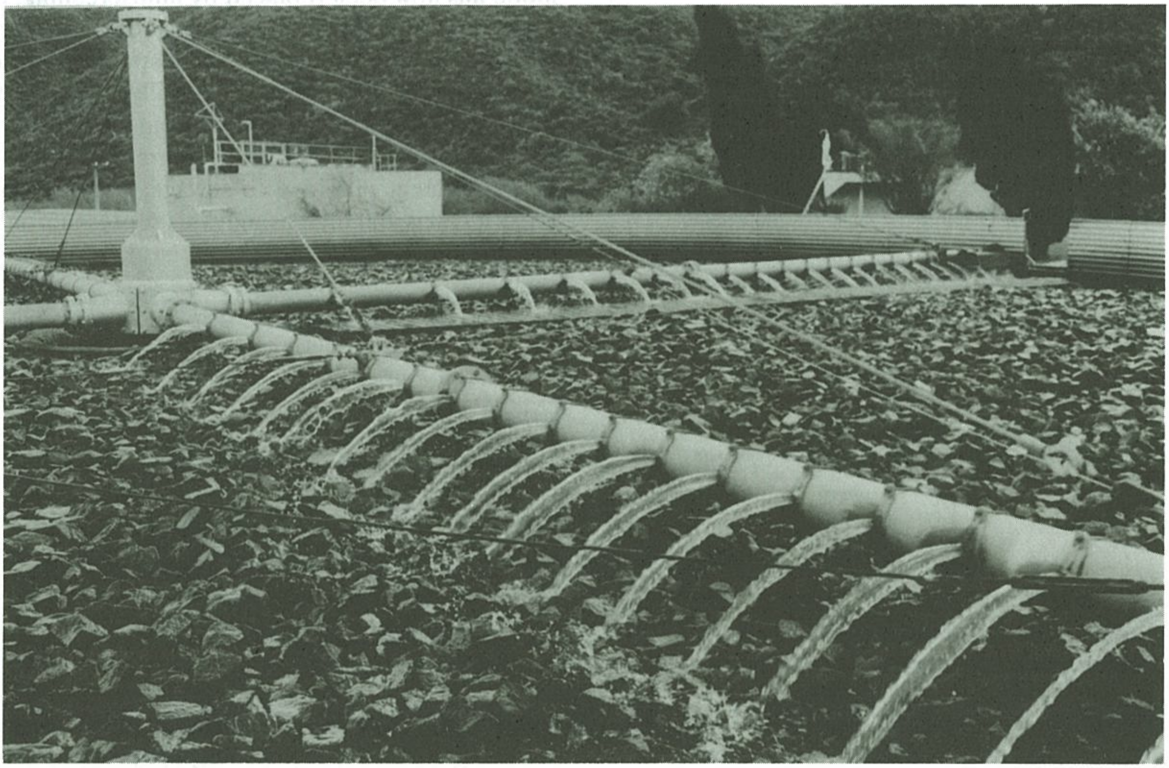

Figure 18.16 Trickling-filter secondary sewage treatment at a plant in Wainuiomata. Effluent is sprayed over rocks where organisms break down the organic substances into stable compounds.

the waste water. Another approach uses what is called activated sludge, where bacteria and other organisms which occur naturally in sewage are used to break down the organic matter.

Tertiary treatment is used to produce "high quality" effluent and tends to use expensive technology. Increasingly tertiary treatment also includes the spraying of high quality effluent onto pasture or forest where natural processes operating in the soil remove the last of the BOD and remaining salts (Davis, 1984).

All sewage contains bacteria, some of which, if given sufficient oxygen, will break down organic material into stable compounds. Apart from being responsible for the odours associated with sewage and sewerage works, these bacteria are not normally harmful. Sewage also contains coliform bacteria, and particularly faecal coliform, which are found in the gut of all warm blooded animals and are excreted in large numbers in faeces. These are not normally considered harmful, and do not reproduce outside the gut, but because they are easy and cheap to measure they are used as indicators of sewage pollution. If a sample of water contains faecal coliform organisms, then it may also contain pathogenic (disease-causing) organisms as well. Whether pathogenic bacteria are present in sewage, and if so their numbers, depends largely the health of the community that is generating the sewage. Sewage from healthy communities will have few pathogenic bacteria (Davis, 1984).

Because of the different types of bacteria which occur in sewage, there are two different sets of reactions which can take place. Aerobic reactions use dissolved oxygen and result in simple end products that are stable, will not harm the environment, and do not smell. Anaerobic reactions, which take place when dissolved oxygen levels are low, lead to smelly organic compounds (e.g. hydrogen sulphide - rotten eggs). Our understanding of the various processes which can occur is now so good that it is possible to build plants within 100 m of urban areas, and this has been done in the United States.

Because the efficiency of treatment is controlled by the length of time sewage remains in the treatment facility, it is important that the plant is not overloaded. In most urban areas overloading occurs because of the difficulty of separating storm runoff from "true" sewage (Figure 18.17). Rain water in a sewer increases the volume of effluent that has to be pumped, treated, and ultimately discharged. It has been estimated that 40% of the cost of constructing a treatment plant is determined by the maximum volume the plant is designed to treat. For example, a plant designed to treat Wellington's peak dry weather sewage flow of 4 m^3/s would cost $40M, while a plant to treat the peak flow during a 20-year return period storm (80 m^3/s) would cost $320M (Kerr, 1991). Also, in most urban areas, toxic industrial chemicals often find their way into the sewers, and these can retard biological activity, and even kill the organisms used in the treatment process.

The Process

An example of a typical secondary treatment facility is the plant at Porirua. This plant treats approximately 23,000 m^3/day (350 l/head/day) of sewage (although it can be three times this amount in wet weather). While predominantly domestic, the sewage does include some waste from industry and local commercial businesses. The bulk of the effluent is water (99.5%) with a third of the remainder being suspended solids and two thirds dissolved solids. The aim of the treatment process is to reduce the suspended solid concentration (from approximately 230 to <30 mg/l), BOD_5 (from between 150-300 to <30 mg/l), and inorganic salts. The inorganic salts come from the high concentrations of ammonia, urea (urine and detergents) and phosphates found in raw sewage. Because plants need both phosphorous and nitrogen to grow, and they can not use one without the other, the plant tries to eliminate the nitrogen to prevent uncontrolled algal growth.

This facility (Figure 18.18) is an extended aeration, activated sludge Carrousel type treatment plant designed to produce high quality secondary treated effluent. The plant uses naturally occurring bacteria in the sewage to break down the organic matter and

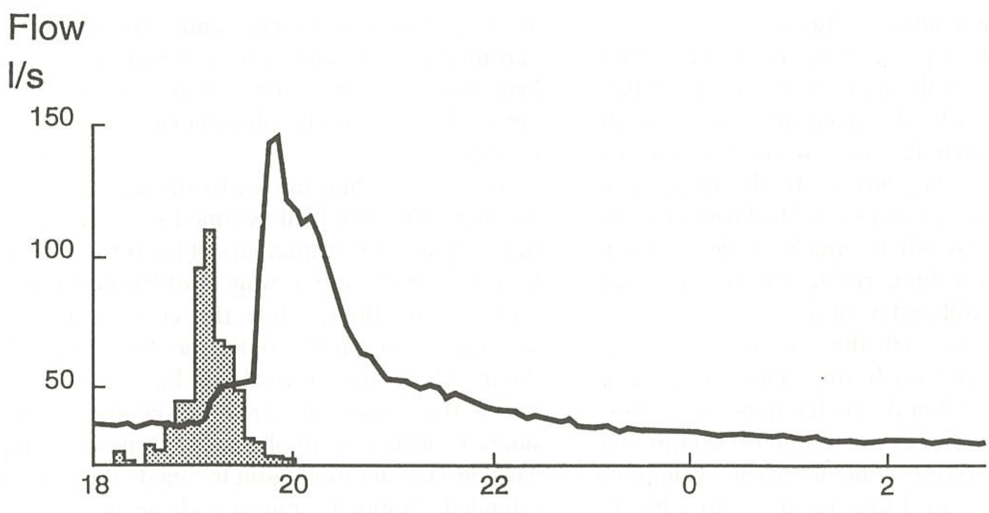

Figure 18.17 Stormwater often enters sewerage systems in urban areas. This hydrograph from a sewer in Wellington shows a very sharp response to rainfall in the "catchment" (After Kerr, 1991).

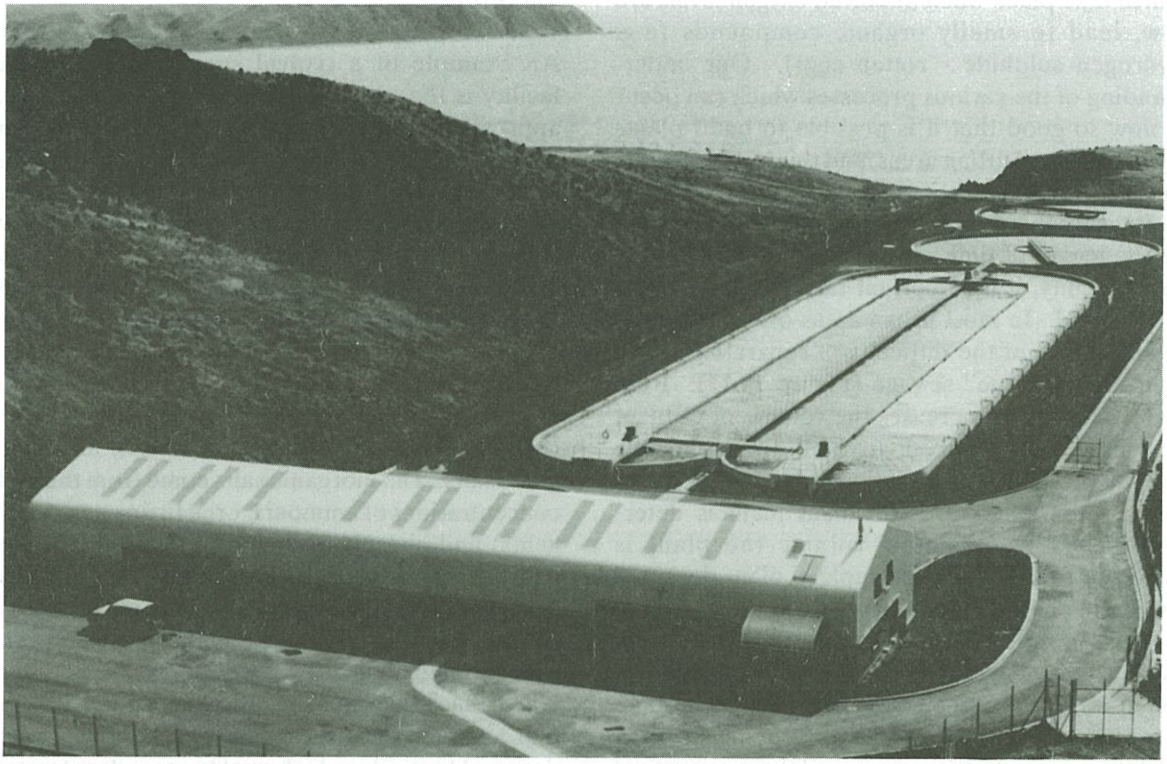

Figure 18.18 The waste water treatment facility at Porirua reduces suspended solid concentrations, BOD, and inorganic salts present before discharging the treated effluent into the sea.

remove nitrogen from the effluent.

The sewage first passes through a milliscreen, with a 0.5 mm mesh diameter, which removes particles larger than this size from the sewage. Even smaller sized particles, and some greases, are removed because they adhere to the larger particles. The screenings, between 3 to 4 tonnes a day, are led to a press where much of the water is removed and the volume reduced before they are taken to the landfill and buried.

After passing through the screens the sewage flows into the aeration basin. This structure is made up of four 4.5 m deep channels, over 100 m long and 9.0 m wide, set out to make a continuous waterway. By carefully controlling the availability of oxygen using air blowers, it is possible to manipulate the kind of bacteria which thrive in each section of the tank. By encouraging the growth of large numbers of bacteria in particular areas, and by placing them under oxygen stress at certain times, the nitrogen in ammonia and urea is broken down to free nitrogen and released to the atmosphere, while the phosphorus is absorbed by the bacteria.

This liquid then moves to the clarifiers, large holding tanks which allow time for the sludge particles (by now large numbers of bacteria etc. which have fed on the raw sewage and formed flocs) to settle to the floor, while the clean water flows through a weir to the outfall as final treated effluent. The sludge of "well fed" bacteria is scraped across the bottom towards the centre collecting sump. From here most of this sludge mass is pumped back into the aeration basin to "feed" the biological extended aeration treatment cycle again.

Because the bacteria reproduce and produce waste, the sludge is a growing mass. Any surplus that is not returned to the aeration basin is pumped to sludge thickeners which release any gas and air bubbles from the bacteria flocs and allows them to settle. The sludge is drawn off and mixed with a polymer to assist the thickening process by coagulation of the solid particles and expulsion of the water from the thickened sludge. A sludge "sausage" is then fed into the belt press where two woven belts gently press the sludge to remove as much water as possible, and the sludge is then dumped at the landfill.

This process removes 99.9% of faecal coliform bacteria and reduces both the suspended solids concentration and BOD_5 to less than 30 mg/l - results of recent tests show levels half of this. This process provides the level of treatment required so that the final treated effluent can be discharged to the sea. Some areas are, however, very sensitive to even subtle inputs of pollutants, and so further purification may be required.

Rotorua has been discharging sewage into its lake since 1882. Even though a secondary treatment plant was completed in 1973, levels of nitrogen and phosphorus remaining in the effluent contributed to excessive algae growth which has deoxygenated the water and caused deterioration of the lake water quality. There was an urgent need therefore to reduce the amount of nitrogen and phosphorus entering the lake. Various options, including pumping the effluent out of the catchment, were investigated but rejected on Maori cultural grounds. It was finally decided that in-plant treatment and land purification would be the most cost-effective means of reducing nitrogen and phosphorus levels.

The final scheme incorporates nutrient stripping via a Bardenpho treatment process (manipulating naturally occurring bacteria) to remove 80% of sewage nitrogen and phosphorus, and then spray irrigation over a 300 ha site planted in pines in Whakarewarewa State Forest (Figure 18.19). As the effluent percolates through the forest litter layer and soil, natural processes "trap" the nitrogen and phosphorus and remove any bacterial organisms that have survived previous treatment.

Figure 18.19 The sensitivity of the aquatic ecosystem at Rotorua has led to the spray irrigation of secondary treated effluent into a pine forest where natural soil processes remove the last of the contaminants.

Studies have shown that denitrification rates (largely through anaerobic means) are particularly high in the riparian zone (along side streams), ranging from 400 to 1200 kg of nitrogen per hectare per year. The efficiency of the system appears to be controlled by nitrate, carbon, and oxygen concentrations with oxygen inhibiting denitrification but the other two factors improving the rate. Riparian soils have low oxygen contents because they are usually water-saturated, oxygen diffusion is limited, and any oxygen which does penetrate is quickly metabolised by aerobic bacteria. They are thus ideally suited to denitrification processes. The carbon which is required as energy to drive the process is readily available from the litter layer and is converted to carbon dioxide.

In a pilot study, the irrigated area developed a lush understorey and the fauna of the forest floor diversified. The number and variety of insects, particularly those associated with the breaking down of litter, also increased significantly. At the same time the volume of litter decreased by up to 50%, indicating that the forest's natural nutrient recycling had accelerated (Forest Research Institute, 1990).

It is believed that this combination of technology and nature will provide an efficient and effective means of purifying Rotorua's waste while recycling the nutrients.

Summary

Human survival has always been intimately tied to the availability of adequate supplies of clean water. The development of an urban society, which concentrates people within small areas, has led to significant impacts on the hydrological cycle - some of which we are only now starting to recognise. Perhaps through having a better understanding of the possible impacts of our activities we can become wiser custodians of this vital resource.

Acknowledgements

Financial assistance to carry out the research for this chapter was provided by the Internal Grants Committee of Victoria University, who also paid for Heather Campbell to draft some of the diagrams. David Winchester also assisted with the diagrams. I would like to thank Fred Carroll and Hadley Bond from the Porirua City Council for the photographs of the 1976 floods and the information relating to waste-water treatment.

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19
Water Resource Management in New Zealand

Andrew Fenemor

Previous chapters have described various components of the hydrological cycle, and the many ways that human activity can affect hydrological processes. Why do we need to know all about hydrology? The main reason is to manage our rivers, streams, lakes, groundwater and wetlands in a wise and ecologically sustainable manner.

Successful water resource management requires knowledge about the occurrence, quality and variability of water resources. We also need to know about demands on the water resource and the community's aspirations for its management. Only then can appropriate measures be taken to manage water in a sustainable way. These measures fall into two categories: regulatory controls on water usage and discharges, and engineering (or structural) solutions like hydroelectricity dams and diversion channels.

In this chapter, we describe how water resources are managed in New Zealand. The emphasis is on the role of regional and unitary councils who are charged with managing water resources under the 1991 Resource Management Act. Major water management issues are illustrated with case studies from throughout New Zealand.

The Need for Water Resource Management in New Zealand

A common perception of New Zealand is that it is a rainy country, the problem being too much water, not too little. Certainly, the early focus of catchment boards was on flood protection and soil conservation. But growing demands for water for urban, industrial and irrigation use led to concern about over-use and water pollution. In some areas, there were already natural water shortages.

In the 1980's, there began a general move away from engineering solutions to water shortage problems (like irrigation schemes) and flood protection, to a more conservative approach of matching demand with availability, emphasising efficiency of water use, and keeping people away from flood waters.

Water management is most needed where demand exceeds availability, both in terms of water quantity and water quality. Demand needs to be defined in its widest sense, incorporating not just consumptive uses like irrigation extractions, but also non-consumptive uses like recreation, fisheries, aesthetic values of streams (often collectively called instream values) and the capacities of waters to assimilate waste discharges.

Water Management Legislation

Soil Conservation and Rivers Control Act 1941

Since 1941, New Zealand has benefitted from a comprehensive regional approach to flood control and soil conservation. The Soil Conservation and Rivers Control Act led to the formation of catchment boards, supported at a national level by the

Soil Conservation and Rivers Control Council. As the name of the Act shows, the early role of catchment boards was soil conservation works and construction of flood protection schemes. This work was a response to the effects of large-scale forest clearance, and a desire to develop floodplains for agriculture and urban growth.

Water and Soil Conservation Act 1967

With the passing of the Water and Soil Conservation Act in 1967, the role of catchment boards was expanded to include regional water board functions. These included the regulation of water use via the granting of water rights , water management, and control of water pollution. A new national policy and advisory body was established, the National Water and Soil Conservation Authority (NWASCA). The Water and Soil Conservation Act promoted the multiple use of water resources, with the provisos that any use of water had to be 'beneficial' and no competing use had a legislated pre-eminence over other uses.

In response to pressure from fisheries and recreational interests the latter proviso was changed to allow for the conservation of highly valued water bodies. The so-called "Wild and Scenic Rivers" amendment in 1981 provided for the protection or preservation, through *water conservation orders*, of rivers, streams or lakes with outstanding natural features. The application for a National Water Conservation Order over the Rakaia River was an enormously expensive legal process, but one which set many of the ground rules for subsequent applications elsewhere.

By the 1980's, regional water boards were developing water management plans for catchments with water shortages or pollution problems. These plans contained guidelines for the granting of water rights , an evolution from the prior ad hoc approach to the issuing of water rights. The preparation of a water management plan provided a good focus for water resource investigations being carried out by regional water boards, to answer the question "What are the issues requiring a policy response in this catchment?". Early water resource investigations identified the extent of existing water management problems in the catchment; more recently, investigations have enabled water managers to deal with **future** resource management problems rather than just reacting to existing problems.

Until recently, water management plans were only informal policy documents, which did not bind regional water boards in the way that district schemes bind district councils. However, the concept of resource management plans has been strengthened in the 1991 Resource Management Act by giving them a statutory basis.

Resource Management Act 1991

The long awaited review of New Zealand's resource management legislation came to fruition in 1991 with the introduction of the Resource Management Act . The management of water resources now shares a common framework with the management of land and air. The key purpose of the Resource Management Act is sustainable management of natural and physical resources, in contrast with the Water and Soil Conservation Act, which supported a multiple-use philosophy.

It is worth reproducing the definition of sustainable management from the Resource Management Act, because this is the goal for future water management in New Zealand:

"Sustainable management means managing the use, development, and protection of natural and physical resources in a way, or at a rate, which enables people and communities to provide for their social, economic, and cultural wellbeing and for their health and safety while -

(a) Sustaining the potential of natural and physical resources (excluding minerals) to meet the reasonably foreseeable needs of future generations; and

(b) Safeguarding the life-supporting capacity of air, water, soil, and ecosystems; and

(c) Avoiding, remedying, or mitigating any adverse effects of activities on the environment."

In a penetrating analysis of the objectives of the Resource Management Act , Fisher (1991) observes that the key word in this definition of sustainable management sustainable management is the word "while". Either the definition means that ecological and environmental sustainability has primacy over the human values, or the intention is to balance use, development and protection of resources and ecological with environmental sustainability. The Planning Tribunal will no doubt resolve this issue in due course.

The Resource Management Act of 1991 provides for many changes in the water management policies of the older legislation (Milne, 1991). The act stresses an integrated approach to the management of all natural resources, with a goal of sustainable management . To do this, it presents a hierarchy of principles, with priority given to environmental concerns, and not just the trading off of benefits and detriments. Water conservation orders can now be applied to any water resource (other than coastal marine areas), including wetlands, groundwaters and geothermal areas.

National environmental standards may be set, and there is more emphasis on monitoring of both the environment and the effects of activities on the environment. The Resource Management Act strengthens enforcement provisions for addressing environmental degradation, and dealing with those who fail to comply with consent conditions .

Regional and unitary councils have been given a central role in water management. The act allows for statutory minimum water quality standards. Maori issues and values receive more recognition, and more public involvement is required when plans are being drafted and consents granted.

The system for granting resource consents for water (previously water rights) has been streamlined, although many of the proven aspects of the Water and Soil legislation have been retained. There are now three types of consents for water: *water permits* (to take, dam or divert water), *discharge permits* (to discharge water or waste where it might pollute water) or *coastal permits* (to take, divert, dam or discharge in the coastal zone).

Administrative procedures for granting consents have been made more flexible, with regional councils able to grant or decline permit applications of minor consequence without public advertising. However, major development proposals now face more rigorous environmental impact assessment requirements. Naturally, the transition to the new system has brought teething troubles, but evolving experience with the Resource Management Act and the writing of plans to guide permit decision-making should, over time, greatly simplify the granting of consents.

Water Management Plans

The major tool for water management - and indeed all resource management under the Resource Management Act - is the management plan. Resource planning involves deciding in advance what to do about existing or potential resource issues, how to do it, and who will do it. It links knowledge to action.

Any resource management plan must satisfy a number of requirements. The desires of resource users, both the community and special interest groups, must be taken into account, and there may be governmental pressures at a national, regional, or local level. Regional and local plans must conform to national guidelines, and be in accordance with government legislation and regulations. Administration of the plan will depend on the structure of the organisations involved in resource management. Technology may hold the answer to some problems, such as waste-water treatment plants, but the most important factor is the resilience and ecological requirements of the resource itself.

Types of Plan

The purpose of plans for water management is to set the rules for managing the water resource through the granting or declining of water and discharge permits .

These plans take many forms under the Resource Management Act (Figure 19.1). *National Policy Statements* contain government policy on matters of national significance. The NZ Coastal Policy Statement is a mandatory statement of this type, prepared by the Department of Conservation.

Regional Policy Statements are mandatory over-

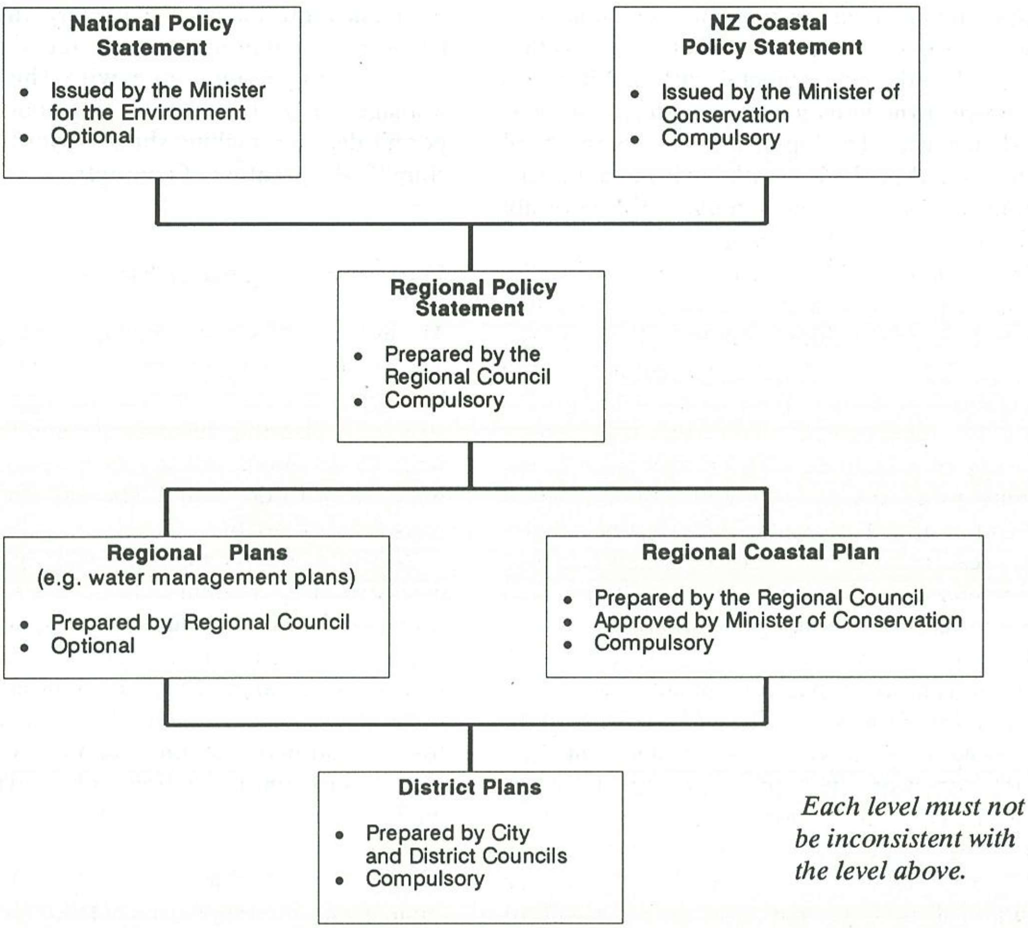

Figure 19.1 Hierarchy of Policy Statements and Plans under the Resource Management Act 1991 (courtesy Canterbury Regional Council).

views of resource management issues in each region, plus policies and methods for integrated management of those resources.

Regional Plans are special-purpose resource management plans, of which water management plans are a good example. The mandatory Regional Coastal Plan for each region is a an example of a regional plan.

District Plans replace the former District Schemes and address territorial functions, particularly in relation to land use and the control of development.

Water Conservation Orders are a special type of national policy instrument which can be instituted to protect or preserve waters which a special tribunal deems to have 'outstanding' values.

These plans have a hierarchy of importance in the Resource Management Act , which requires that no plan can be inconsistent with those higher in the hierarchy. Regional water management plans therefore must not conflict with the Regional Policy Statement, any National Policy Statement, or applicable Water Conservation Order.

The first three types of plan are those most likely to address water management issues. Given the history of water management planning in regional water boards, it seems likely that water management plans for specific catchments and issues will continue to be developed as separate regional plans, but with the statutory force given them by the Resource Management Act.

The Planning Process

A plan is only needed where there is a problem, or one is likely to develop. Plans can easily become huge unwieldy documents which mix important issues with trivial ones, and swamp the reader with unnecessary information. It is thus important to separate resource information from plan policies and rules. Plans which are not easily read and understood will not be complied with. The corollary, however, is that policies and rules require sufficient justification for the user to understand the reasons for those policies and rules.

Figure 19.2 shows the main components of the water management planning process. There must first be issues which need to be addressed by a plan; these will define the goal and objectives of the plan. Our experience is that in water management, many issues are only revealed through water resource investigations. When a problem has already occurred, such as seawater intrusion into an aquifer or a fish kill in a river due to excessive abstraction of water, it is more difficult to solve the problem.

The choice of policies in a plan will be influenced by the existing users of the resource. In some instances there are over-riding national interests in a particular resource and these special values may be protected by measures set out in a Water Conservation Order. For example, the Rakaia River Conservation Order protects instream habitat and flows, to allow for salmon migration and the preservation of the habitat of rare and endangered birds.

The Resource Management Act requires planners to consider alternatives when adopting policies and rules, and to evaluate the benefits and costs of each alternative. Planning options will usually cover the range from "do nothing" to a tightly constrained regulatory approach. They may include operational responses such as implementation of a water augmentation scheme.

The most important aspect of the planning process under the Resource Management Act is public consultation. Water users and affected parties must see the plan as theirs. Finally, the success of a water management plan is judged from its implementation - it needs to be monitored to ensure it achieved what was intended. This monitoring, plus further information, will enable the plan to be corrected or refined over time.

The Role of Hydrology in Water Management

If we define hydrology in its broadest sense, to cover the flow and quality of water in the natural environment, it becomes obvious that hydrology provides the information base for water management.

Knowledge about the water resource you are managing is vital. New Zealand's economic decline is putting pressure on funding of water resource data collection, along with other expenditures, thus requiring hydrologists and water managers to justify their data collection networks. It has resulted in more targeted investigations, but these are usually aimed only at answering short-term problems. It has also reduced access to resource data, because of supposed confidentiality or high charges to make some data available.

Among the most important management problems are floods and droughts ; long-term hydrological records are needed to determine the frequency of occurrence of these events. Long-term records should continue to be collected, both for these reasons and because many of the uses of these data have yet to be identified (eg. effects of climate change on river flows).

Water resource investigations are needed to address many key areas:

Water Availability - The amount of water available from a river or aquifer varies with time. In rivers, water availability is dependent on flow alone, but in lakes and aquifers, storage also is important. In many gravel bed rivers, substantial amounts of water leave the river and enter the shallow unconfined aquifer systems, e.g. the Ashburton River in South Canterbury.

In geothermal systems such as at Rotorua, the

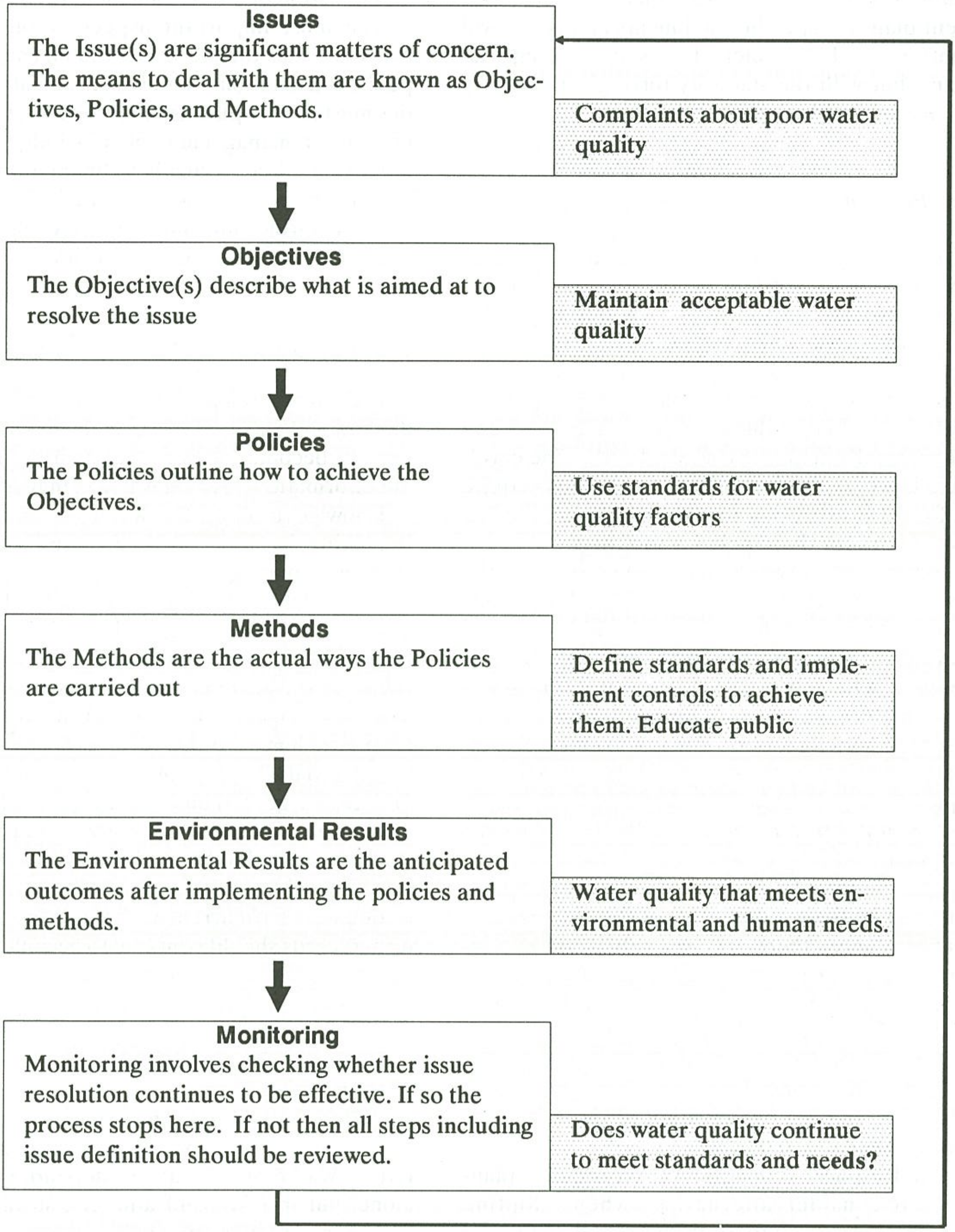

Figure 19.2 Water Management Planning Process using Water Quality as an Example (courtesy Canterbury Regional Council).

additional parameters temperature and pressure affect availability of hot water and behaviour of geysers and mud pools.

Water Quality and Contamination - To determine the degree of pollution of water, it is necessary to know the natural water quality. Pollution is most easily reduced at its source. Contamination can come from a single point source, such as a sewer outfall, or from a diffuse source such as runoff containing fertilizers leached from pastures. The measures needed to reduce contamination will depend on the nature of the source, and the desired level of quality of the water.

Instream Values - This term encompasses the full range of natural, aesthetic and recreational values of a water body, some of which are social and cultural values foreign to hydrologists. Examples include flow needs for canoeing, scenic appreciation, fish habitat, and provision for fish passage.

Advances in technology have greatly increased the accuracy and ease of collecting hydrological data. Computers and telemetered data collection now allow a more immediate response to floods and droughts. In the future, development of sensors for a wider range of water quality parameters promises to improve responses to water-related environmental problems. Technology has enabled huge amounts of hydrological data to be collected and stored. The challenge is to analyze and understand the hydrological processes it represents.

Recent Trends in Water Management

Integrated Environmental Management

The Resource Management Act adopts an ecosystem approach to resource management. This is nothing new - the National Water and Soil Conservation Authority encouraged the integrated management of land and water, because land uses can so dramatically affect water quality and flows. However, legislating this approach will strengthen recognition of links within ecosystems. One benefit will be a closer relationship between district plans and regional plans. For example, controls on exotic afforestation may be introduced in the Tasman District Council's District Plan to prevent excessive reductions in stream flows and groundwater recharge in the Moutere Valley near Nelson.

Increased Public Consultation

Water management plans drafted by regional water boards were early examples of consultation with resource users. The Resource Management Act now requires local and central government to consult with agencies, resource users and the public when preparing plans. This is to ensure that all those affected by the plan are aware of the proposal and can express their opinions so that the final plan takes their wishes into account. This approach allows mediation and negotiation of policies rather than confrontation and the resulting expensive litigation.

Electricorp negotiated with potential objectors prior to the lodging of applications to renew resource consents for its hydroelectric dams on the Waitaki River. Although compromises on residual flows and enhancement of fisheries and wildlife were necessary to reach agreement, more money was able to be spent on productive work rather than expensive legal arguments (M France, ECNZ, pers. comm.).

Similar success has occurred in consultations between the Manawatu-Wanganui Regional Council and Oroua River water users and dischargers. Voluntary agreement was reached on providing a core allocation of water which would be always available to irrigators. Agreement was also reached on a timetable for improving the quality of meat processing and sewerage discharges into the river. Details of these agreements were incorporated as conditions on water and discharge permits granted to water users (B Cowie, Manawatu Wanganui Regional Council, pers. comm.).

Other examples of effective consultation are the water user committees set up by regional councils in many parts of New Zealand. These user groups provide the contact between the regional council and water users. In catchments where there is conflict between small numbers of consumptive water users, user committees may have delegated power to allocate the water among users during times of shortage. Where there are minimum flows, monitoring requirements, or broader conflicts, the regional council usually remains the regulator but liaises with user committees over action such as rationing of water usage.

Economic Instruments

Although concepts such as "polluter pays" and tradeable water permits were widely discussed during the drafting of the Resource Management Act, there is not yet extensive use of these methods. One legislated economic instrument is the ability to transfer a water permit from place to place (not just to a successor landowner, as previously) if specifically allowed in a regional plan. Thus, a landowner could sell his permit to a neighbour, and the council would register that transfer.

Economic instruments have not been widely used for resource allocation to date because the effects of resource use on third parties are often difficult to address. For example, trading of permits to take water from a major river may require rules on the locations and quantities for abstraction to maintain a viable trout fishery in that river.

The Maruia Society has become a strong advocate of the use of economic instruments as part of its "Green Economy" philosophy, even to the extent of supporting the idea of tradeable discharge permits (Salmon, 1991). However, initially we are more likely to see trade in permits to take water because water allocation is simpler than water quality management.

Where demand exceeds supply, trade in water permits may be workable (Fenemor, 1991). The tradeable water permit system must, however, lead to a more efficient water management system than a fully regulatory approach. Detailed knowledge of the availability and dynamics of the water resource is needed, so that constraints to trade are well defined, and there must be adequate protection of instream, Maori and other values built into those constraints.

There are likely to be further developments in this area of resource economics in the near future, particularly as demand on some water resources increases.

Recognition of Maori Interests

The Purpose and Principles of the Resource Management Act (sections 5 - 8) require resource managers to explicitly take into account Maori values. The Act gives national importance to *"the relationship of Maori and their culture and traditions with their ancestral lands, water, sites, waahi tapu, and other taonga"*. Resource managers are also to have regard to *kaitiakitanga*, i.e. the concept of guardianship of natural resources. The principles of the Treaty of Waitangi are also to be taken into account.

Maori regard water as a taonga or treasure left by ancestors for the life-sustaining use of their descendants. The descendants are in turn charged with stewardship duties, resource management or rangitiratanga, and kaitiakitanga, to ensure these treasures are passed on in as good a state to those following. These concepts fit well with the intent of the Resource Management Act to allow for the needs of future generations. Tribes or hapu pride themselves in sustaining valued resources. Because the marae depended on these resources, the resources assume great mana (Taylor and Patrick, 1987).

Water is considered by Maori to possess a life force or mauri, and to have a spirit or wairua in relation to the quality and use of that water. For Maori, the physical and spiritual realms are linked. Thus, the taking of large proportions of a river's flow could be considered to be dismembering the body of that river. Likewise, the discharging of effluent into waters, whether used as a source of food (kaimoana) or not, may be desecrating the mauri of that water, even when science indicates that the discharge is "acceptable".

Long before the Resource Management Act became law, water managers in those areas of New Zealand with a large Maori population have been under pressure to take Maori concerns into account, especially when granting water rights. People involved in water management must develop their knowledge of local Maori iwi, their values, and protocol on marae. Maori views on resource issues are often best gauged from a hui or meeting on the marae. Water managers need to be active in obtaining Maori views.

Local bodies in Northland, for example, have responded positively to Maori opposition to effluent discharges into water, by extensive development of wetland sewage treatment systems. Constructed marsh systems are being used for tertiary treatment of effluent, and are reducing bacterial and nutrient discharges to streams and the coast (L.Parker, Northland Regional Council, pers. comm.).

Increased Compliance Monitoring

There is recognition in the Resource Management Act that resource consents must be monitored for compliance with their conditions, or they are open to abuse. Routine monitoring of compliance was initiated by regional water boards in the early 1980's, following an amendment to the Water and Soil Conservation Act which allowed boards to charge consent holders. Most regional councils now have annual monitoring programmes for their major discharge consents . Initial opposition to charges for monitoring has passed, and many larger enterprises now view a good environmental record as enhancing their image. Non-compliance problems usually arise in cash-strapped industries.

In areas where water resources are fully allocated or inadequate, permits to take water are monitored by metering. Water metering is useful not only for checking compliance with allocations, but is essential for the user and the regional council when allocations are rationed. Many irrigators have found water meters useful for their irrigation scheduling. Water meters also provide valuable usage data to water managers for investigating the impact of usage on the resource, so that allocations can be refined.

Water Conservation Orders

The Resource Management Act provides for Water Conservation Orders to preserve waters in their natural state, or to protect waters already partially modified.

Between 1981 and 1992, twenty-three water conservation orders were applied for but only thirteen have been finalised and granted.

Water conservation orders frequently have a "no dams" clause to safeguard outstanding trout and salmon fisheries, e.g. Rakaia , Mataura and Motueka Rivers . This provision also allows for the preservation in their natural state of "wild and scenic" sections of rivers which have outstanding values for recreational uses such as canoeing, rafting and jet-boating. e.g. Motu , Buller and Mohaka Rivers.

Conservation orders can be used to set rules or guidelines to protect water quality in water bodies. The very high natural water quality of Lake Wakatipu and its inflowing rivers (Routeburn, Caples, Greenstone etc) has been recognised in the draft order for the Kawarau River catchment. In the Queenstown area, high water quality and the wild and scenic rivers underpin a large part of the tourism industry. In 1991, ticket sales alone for water-based activities were worth $25.4 million.

Under the Resource Management Act, a water conservation order may also apply to wetlands . Part of the Kawarau draft order provides for the "preservation in their natural state" of the Nokomai string-bogs in the headwaters of the Nevis River , a quite remarkable and fragile landscape feature. These are bogs, quite unique in New Zealand, which would normally be found in sub-arctic areas like far northern Canada or Siberia.

Surface Water Management Issues

Planning for Extremes

Adequate supplies of water from stream flow or lake storage may not always be available, especially during dry summers. There is thus a need to plan for these extremes. An example is the winter drought of 1992 which affected hydroelectric power generation in the South Island, when exceptionally low inflows over a seven month period severely restricted the water available.

Another good example is the Opihi River in South Canterbury, where a regional council water management plan allocates water between irrigation and instream users , with a threshold flow below which irrigators must cease pumping (South Canterbury Catchment Board and Regional Water Board, 1984). Good hydrological flow data are needed to set the threshold flows in this type of plan.

Lack of Information

Information may be lacking on the water resource and the impacts on it, especially during the extremes of drought or flood . For water related projects, impacts are sometimes estimated, then the estimates are checked by monitoring when the project is implemented. For example, the Tekapo River flow has been greatly reduced by diversion for hydro-electric power generation. Monitoring showed that the residual river was actually a better trout habitat, but was being adversely affected by flood releases of water from Lake Tekapo. Studies commissioned by Electricorp in 1989-90, as part of the process of obtaining resource consents for the Upper Waitaki system, showed that the flood releases from Tekapo dam could be controlled to mirror natural flood flows. This change to operating procedures should reduce the undesirable environmental impacts in the Tekapo River.

Conflicts between Water Uses

One type of use can affect others, for example abstraction for hydropower generation may conflict with the needs of canoeists. In the Wairoa River near Tauranga for example, provision has been made for hydro-electric stations to restore natural river flows on some weekends for river rafting and canoeing. Similar arrangements allow canoeing races to take place downstream of the Mangahao power station, near Palmerston North.

Balancing Consumptive Demands and Instream Values

The minimum flows needed for recreation, aesthetic appeal, and fisheries are very difficult to determine and are often based on subjective judgement. In the case of fish habitat, some of this subjectivity has been removed using hydraulic modelling techniques such as Instream Flow Incremental Methodology (IFIM) (Jowett, 1982), in which habitat variables like water temperature, velocity, depth, and substrate are related to flow. Jowett (1991) extended this approach by developing a hydraulic model of "weighted usable area" for brown and rainbow trout.

For recreational uses such as canoeing, minimum flows and depths can be determined by relating canoeists' assessments of their canoeing "experience" to the flow conditions. Of course, many different instream values exist for a river, and there may be varied flow requirements for each.

Interpreting Sustainable Management

The legal requirement for sustainable management is central to the Resource Management Act This requirement does not simply equate with balancing water demand and availability. Nor is it met by balancing competing uses, for example abstractive irrigation, habitat requirements and the ability of rivers to assimilate polluted discharges. This multiple use concept was advanced by the former Water and Soil Conservation Act.

Sustainable management can be interpreted as setting a "biophysical bottom line", meaning that sufficient water must be left in rivers, streams, lakes and wetlands to maintain the aquatic habitat, invertebrates, fish and bird life and also provide for aesthetic values and recreational uses of the water body. This does not preclude abstractive use such as damming and diversion, but it requires that environmental "values" or "uses" must be taken into account.

Allocations Differ from Actual Usage

The amount of water allocated in resource consents may differ considerably from the quantities actually used. Simply adding together the quantities allocated for extraction on water permits gives a misleading picture of demand, because water is not extracted all at once, and may not reach the maximum allocated, even in a major drought.

In Nelson, for example, metered weekly irrigation water usage on the Waimea Plains was never more than 68% of total allocations during the 1-in-10 year drought of 1989-90 (Nelson-Marlborough Regional Council, 1991).

Similar results were obtained for the Pareora River in South Canterbury; in ordinary summers, water use may be as low as 30% of the allocation. (J Waugh, pers. comm.). Matching usage with allocation requires increasingly detailed management controls, which may not be justifiable from a cost-benefit point of view.

Water Quality Standards

Water quality defined in its widest sense includes chemistry, biological quality, physical characteristics such as colour, and its acceptability for use, including aesthetic acceptability.

Suitable standards for water quality can be defined to assist in water quality management. The Resource Management Act differs in this respect from the former Water and Soil Conservation Act, in that it provides descriptive water quality standards defined by the use of the water, rather than specifying quantitative standards. Regional councils can determine what quantitative measurements they will use to check compliance with standards.

National guidelines have been issued by the Ministry for the Environment on the optical quality of water and the control of undesirable biological growths in water. Microbiological water quality standards for recreational and shellfish gathering waters have been issued by the Department of Health.

One problem with setting minimum standards is that they become the bottom line to which the water quality may decline in the long term.

Pollution from Diffuse Runoff

The control of diffuse runoff of fertiliser, animal wastes, pesticides, contaminated stormwater, and sediment into water bodies is a difficult issue. It is best addressed by integrating land and water management practices. Regional councils have tackled this in two ways. Submissions may be made on district plans to ensure that potentially polluting land uses such as factory farms or hazardous waste stores are either sited well away from water, or operated so that the risk of pollution is minimised. Secondly, regional land or water management controls may be imposed, for example, to minimise erosion or the discharge of highly contaminated stormwater.

The Auckland Regional Council has a programme to make city residents aware of the contaminants which can enter stormwater . One innovative part of their public relations campaign has been to paint stylised fish beside each stormwater intake on the streets, to remind people that fish live in the estuaries where much of the stormwater ends up. The Regional Council is also constructing urban lakes to act as retention and settling basins for some of the pollutants carried by stormwater, before it is discharged to the sea or to streams. The heaviest load of pollutants is picked up at the beginning of a storm, so this is an effective way to reduce the impact of stormwater discharges. Retention and settling have little effect, however, on dissolved contaminants.

Vegetated stream banks are valuable as interceptors of runoff, and act as filters for sediment and nutrients entering watercourses as diffuse runoff. Much more could be done by regional councils to encourage landowners to plant riparian vegetation. The Department of Conservation has commissioned research by the Water Quality Centre, Hamilton, to produce guidelines on the use of vegetated riparian strips or zones.

Water Quality Monitoring

Water quality monitoring is both complex and costly. Waters may be polluted by a myriad of contaminants; to find out whether water is polluted, you have to know what pollutant to look for. Equally difficult is the problem of determining when and where pollutants entered the water body.

This complexity underlines the need for research to understand the hydrology of rivers, streams and lakes so that useful, targeted water quality investigations can be carried out.

Contingency Planning

Contingency plans are needed to avoid major spills, because these usually cause greatest localised damage. The Water and Soil Conservation Act seemed ambivalent about contingency planning as a legal requirement, but the wider definition of environmental "effects" in the Resource Management Act now gives a stronger mandate for requiring contingency plans from potential polluters, for example at timber treatment plants where highly toxic chemicals are in routine use.

Public Awareness of Pollution

While most people would profess a concern for the environment, many do not realise the impact that their activities may have. Discharges into urban stormwater systems are a good example.

Change is needed to some people's perceptions that dumping of refuse into or beside a watercourse is an acceptable means of disposal. Because the pollutant is carried away quickly, the perpetrators are hard to catch. Educational campaigns by councils and environmental groups are raising awareness of water pollution.

Dissemination of scientific advances to water managers and the general public also helps to raise awareness and solve existing problems. An example is the initiative of Landcare Rotorua (formerly Forest Research Institute) in attracting members to the "Land Treatment Collective" , which holds seminars and field days on the land treatment of wastes.

Groundwater Management Issues

Many constraints will limit the quantity of water able to be extracted from a particular aquifer system. Groundwater quality can likewise limit the uses of that water.

Seawater Intrusion

If pumping draws down the piezometric head in an aquifer below sea level long enough, seawater will enter the aquifer. In confined aquifers, seawater intrusion can be disastrous because it takes a long time for the seawater to be flushed out of the aquifer after pumping has stopped. In unconfined aquifers, the higher groundwater velocities will flush out the salt more quickly, but in both cases, pumps have to be shut down to allow this intrusion to reverse.

As an example, seawater intruded 600 metres inland into the Motueka Gravel Unconfined Aquifer at Lower Moutere, near Nelson, in March 1990, forcing irrigators, whose pumping caused the problem, to cut their water use. Measurements of groundwater salinity showed that the groundwater returned to potable quality within two months of the end of the irrigation season. A privately operated water supply scheme now supplies irrigation and domestic water from an alternative source to the affected area.

'Mining' of Groundwater

Overallocation of groundwater may cause an unacceptable long-term decline in the water table, sometimes called 'mining'. In New Zealand, groundwater mining is a potential problem in some of the small basalt aquifers of the North Island, which have low recharge rates.

Aquifers in alluvial gravels are less vulnerable because they are recharged from rainfall on the surface of the plains and from the rivers crossing the plains. In Canterbury, leakage of water from large areas of border-dyke irrigation is also a major source of recharge.

In the Rotorua Geothermal Field, reduced thermal activity at Whakarewarewa has been caused by excessive utilisation of hot water and steam. This led to the closure of many bores by the Ministry of Energy in 1987-88 to reduce the drawoff. The management plan for the field sets priorities for use ranging from public mineral baths down to domestic mineral baths and general space heating (Bay of Plenty Catchment Board, 1988).

Localised Decline Water Table

Unacceptable localised drawdowns in the water table often occur when pumped wells are located too close to each other. Interference drawdowns between pumped wells must be considered when granting permits to take groundwater, and this requires data from pumping tests describing the permeability (or transmissivity) and storage characteristics of the aquifer. Water management plans often specify minimum spacings for production bores.

One difficult management issue is the loss of supply to shallow domestic wells caused by pumping from deeper wells. Protecting domestic groundwater availability may severely restrict utilisation of the available resource. In West Melton, near Christchurch, irrigation pumpage is rationed to protect domestic groundwater users (J Talbot, Canterbury Regional Council, pers comm.), while on the Motueka Plains, domestic wells must penetrate the full depth of the aquifer before their supplies are protected (Tasman District Council, 1992).

Aquifer Subsidence

Excessive pumping from thick, compressible aquifers , especially those with a high proportion of clay, reduces the pore-water pressure within the aquifer, and this can result in irreversible compaction of the aquifer matrix. Slumping of the ground surface can result, causing damage to buildings and services. Severe aquifer subsidence has occurred, for example, in Mexico City, Florida and Venice.

Stream-Aquifer Interaction

Where rivers, streams and lakes lose water into aquifers , extra pumping from the aquifer can induce additional recharge. This may be acceptable, provided minimum stream flows are not compromised, and provided the surface water quality is suitable for recharge. Connected surface water and groundwater systems should therefore be managed as a single hydrological entity.

A good example of this integrated approach is the water management plan for the Waimea Plains rivers and aquifers near Nelson (Nelson-Marlborough Regional Council, 1991). A major concern in managing this water resource is the maintenance of a minimum flow in the Waimea River during dry summers. Other concerns are the prevention of seawater intrusion into the confined and unconfined aquifers, and the need to augment the summer resource, given that the water is now fully allocated.

Groundwater Resource Information

Successful water management relies on an adequate knowledge of the hydrological system. For a groundwater system, information is needed on the geographical boundaries of aquifers, recharge and discharge mechanisms, hydraulic characteristics and groundwater quality. Assessments of the impacts of pumping individual wells can be made by simple calculations, but cumulative impacts of pumping multiple wells are best determined using computer modelling. Ongoing monitoring of groundwater level and usage (and in the case of geothermal systems , temperature and pressure) is essential to determine the effectiveness of management policies.

For the Waimea Plains aquifer system a three-dimensional computer model was developed to determine the safe yield of the system (Fenemor, 1988). Model development was the impetus for interpreting a large amount of hydrological and geological data collected during the previous fifteen years (Dicker et al., 1992). The Waimea Basin is now one of the best understood geohydrological systems in New Zealand, and one of the most heavily utilised.

Groundwater Contamination

Aquifer contamination is an insidious process which can remain undetected for a long time. When discovered, remediation is a complex, expensive process. Groundwater contamination is currently one of the major environmental issues in the United States, and is becoming an increasing problem in New Zealand. Water resource managers can address the issue in the same ways as described already for surface waters. Particularly important are recognition of diffuse sources of pollution such as fertilizer leaching, and the need for contingency plans to avoid spills of contaminants. Current groundwater contamination issues in New Zealand include hydrocarbon pollution from underground storage tanks, leaching of pesticides and other contaminants from dumps, and nitrate pollution from stock, dairy waste and fertilizers.

Lessons for Groundwater Managers

Among the lessons learnt about groundwater management are the need to manage the whole system as one.

Determining the sustainable yield of aquifers is a challenge posed by the Resource Management Act . This will depend on the pattern of extraction and the limitations for groundwater management such as seawater intrusion . A common assumption when evaluating aquifer yield is that the sustainable yield is equivalent to the natural mean annual recharge. This may or may not be correct; if pumping from an aquifer induces increased recharge, then pumping more than the natural recharge can still be sustainable.

However, a run of dry years, such as that which occurred in Canterbury in the early 1980's can lead to progressive lowering of regional aquifer levels. In this situation it may not be wise to continue extracting water at a rate equivalent to the natural mean annual recharge. In the Canterbury drought a few deep wells went dry.

A third lesson is the need for water resource managers to understand the dynamics of groundwater systems before over extraction occurs.

Acknowledgements

John Waugh contributed useful information on water conservation orders, and with Eileen McSaveney assisted greatly with editorial revisions. My thanks to those regional council staff who provided case study material for inclusion in this chapter.

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20
Water and Slope Stability

J A McConchie

Introduction

New Zealand's landscape is characterised by widespread slope instability. While most slope failures are triggered by the interaction of water and the slope material, many of these sites are predisposed to instability because of New Zealand's dynamic tectonic and climatic history. The length of time it takes the landscape to adjust to changes in conditions has led to many slopes being out of equilibrium with contemporary processes. Many slopes are thus preconditioned for landsliding and require only a small change in one factor to make them fail. Final failure can be caused by either an increase in the shearing force (the force tending to make the material slide down slope) or a decrease in the material's shear strength (the properties holding the material onto the slope). Water content, because it varies rapidly, is the most common trigger of slope failure. If we understand the role of water in slope stability we can manage the landscape to minimise the risk.

New Zealand's landscape is generally steep, geologically young, tectonically active, and developed on soft or highly jointed and faulted rock. The axial ranges are aligned across the path of the prevailing westerly weather pattern, producing heavy rainfall. There are thus many instances of slope failure under natural conditions. Widespread grazing of slopes (up to 35°) which have been cleared of forest, and the urbanisation of progressively steeper slopes, have however certainly increased the incidence of landslipping over the last 150 years. Given this high incidence of slope failure there have been surprisingly few deaths recorded. Hawley (1984) argues that *"those citizens who do not either venture into the mountainlands, or work in service trenches, are exposed to a risk of less than one chance in ten million of being killed by a landslide in any year. The same citizens are exposed to a risk about 2000 times greater than this of being killed on the roads."* This apparently low rate can be explained largely by New Zealand's low population density, and the fact that the majority of landslides are either relatively small or slow moving.

Although the number of fatalities from landslipping is low, the cost is not. Perhaps the most dramatic illustration of the costs of landslides relates to the Clyde Power Project. Investigation and stabilisation of a series of ancient landslides upstream of the dam required approximately $400M, 14.5 km of tunnels, 60 km of surface drilling, and 78 km of drainage holes drilled from within the tunnels. Five million cubic metres of buttressing material was placed at the base of the slide areas to improve stability.

The average annual cost of landslips in the pastoral hill country is estimated to be in excess of $20 million. This, however, includes only the immediate costs and not the ongoing cost of lost productivity (Hawley, 1984). New Zealand's hill country farmers have had to learn to live with the disruption caused by buried fences, loss of pasture, debris-filled stock dams, and blocked tracks and roads caused by landslipping. Between 1970 and

1981, only 1972 did not have a landslip-inducing storm severe enough to be recorded in the Head Office files of the Water and Soil Division, Ministry of Works and Development. The 26 storms during this period were widely distributed across the country. There would also have been an unknown number of storms, smaller in magnitude, less extensive, or too unimportant economically to have been subject to reports (Eyles and Eyles, 1981).

The cost of these storms varies widely, but two examples illustrate their impact. Wairarapa experienced an exceptionally wet winter in 1977 which resulted in a large number of shallow soil slips. Farmers in the 1400 km^2 area most severely affected suffered an immediate loss in annual income of approximately $1M. For the one hundred farmers who lost 15% or more of their income, the average loss was 30%. The high cost of this localised storm was caused by the large number of stock that died through lack of feed, because debris from the slips covered the more productive lower pastures. This initial cost, however, was not the total impact. While slip scars recover their grass productivity quite rapidly over the first 20 years, to 70-80% of that of neighbouring uneroded soil, recovery tends to remain at about this level. No greater recovery was found on scars up to 80 years old. On these slopes productivity has been reduced by 20% from repeated episodes of landslipping (Hawley, 1984).

More recently, Cyclone Bola (1988) devastated a large tract of the North Island. The Government provided $56M in direct subsidies to assist agriculture and $120M in total. To this must be added the amount paid by insurance companies, approximately $37M. This reflects only the immediate cost of an event which has left a lasting impact on the people and productivity of the area.

Storms resulting in landslips in urban areas can be more expensive, at least initially. A 24-hour storm which affected a narrow band of Wellington's hilly suburbs in 1976 caused $20M of damage from flooding and landslip (Bishop, 1976). In 1991 the Earthquake and War Damage Commission paid out on 406 claims with a total value of over $1.7M (Figure 21.1).

Landslides therefore impose a high cost on society, but through understanding the processes causing failure the risk can be minimised.

Concepts of strength

The relative stability of a slope comes from the balance between the strength of the material and the stresses acting to make it move down slope. Any material that is on a slope is inherently unstable, although not all slopes will fail. This is because the strength of most materials is sufficient to resist the forces imposed by gravity. In examining the stability of slopes it is important to understand the nature of this shear strength and the factors that can reduce it to a critical level.

Sources of Strength

The strength of materials forming slopes depends on the type of particles that are present and the way they are packed together. This strength comes from either the friction or the attraction between the particles. If you imagine a typical soil as being composed of spherical particles, rather like ball bearings, then it gets its strength largely from friction between the particles. This friction is of two kinds: planar friction - the friction between one particle surface and another, and interlocking friction - the strength that comes from one particle having to "climb over its neighbours" to shift position. In material forming slopes these two sources of friction, and therefore strength, are usually inseparable, because it is impossible to determine the individual component contributions accurately. Frictional strength increases if the force or weight pushing the two surfaces together increases. For example, it is much harder to push a block of wood across a table if there is a force pushing down on top. The forces pushing the particles together are therefore very important in controlling the strength of a material, and are termed the normal stress.

If we now imagine that we have "magnetised" our ball bearings, then they have even more strength. Not only do they have their frictional strength as before, but they now have an extra "grip" because of the electromagnetic attraction between the bearings. This is known as cohesion and is a measure of the "stickiness" of the particles. Cohesion not only increases the overall strength of

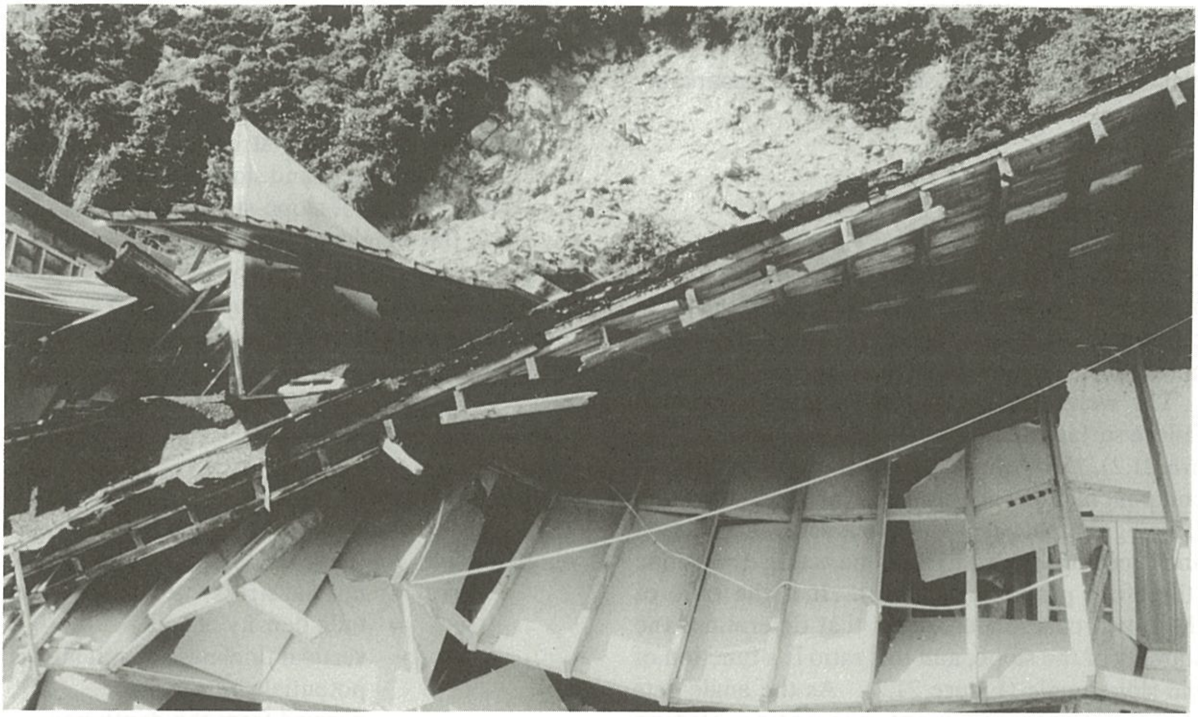

Figure 20.1 Even relatively small, shallow, surficial landslides can be destructive. This house was demolished by a landslide from the coastal cliff at Eastbourne near Wellington.

the material, but it is independent of the normal stress i.e., it remains constant irrespective of the amount of force pushing the surfaces together. Cohesive strength normally includes all strength that is not attributed to friction e.g., suction, electrochemical attraction, and the effect of vegetation roots. The simplest interpretation of strength viewed in this way was provided by Coulomb in 1776 who presented a model of strength where:

$$s = c + \sigma . tan\phi \qquad (21.1)$$

where s = strength
c = cohesion
σ = normal stress
ϕ = angle of internal friction

This is simply a regression line that shows the relationship between strength, cohesion, normal stress and friction.

Theoretically, there are two classes of materials: cohesive, and cohesionless. In reality, however, many materials have intermediate characteristics of strength and behaviour, or change their behaviour depending on moisture content and the stresses present. For example, if you hold a candle by both ends and apply a slight downward pressure the candle will through time, bend perhaps even forming a complete circle. If the same force is applied suddenly, the candle snaps rather than bends. While its strength has remained constant its behaviour has changed markedly.

Obviously the strength of most slope materials is considerably more complicated than the above analogy and model suggest. For example, strength has been shown to be a function of void ratio, friction, cohesion, composition, history, temperature, strain, strain rate and soil structure. In most situations, however, the above simple model explains the behaviour of slope material remarkably well.

Normal Stress

A critical factor controlling the strength of a material to resist any downslope stress is the force pushing the particles together. This force comes mainly from the weight of the material.

The weight of the material above a potential failure surface acts in the direction of gravity (Figure 21.2). However, this resultant force has two components: one acting at right angles to the slope (forcing the particles together, increasing friction and therefore strength) and one acting downslope (therefore promoting failure). It is the ratio of these two component forces that determines the stability of the slope, and this ratio is a function of the slope angle (Figure 21.2). As the angle gets steeper the component promoting sliding lengthens and the component promoting "grip" shortens. On slope angles commonly found in the field, the weight contributes more to the strength of the material than to its potential to move. The weight is therefore very important in maintaining stability, because it pushes the particles together and increases friction.

Using this simple model, it is possible to calculate the magnitude of the stresses acting in both directions, if the weight and slope angle are known. This is the essence of slope-stability analysis. On any slope the weight of the regolith or rock is a function of its bulk density (weight per unit volume), thickness, and the slope angle. The stresses acting in each direction can be calculated from the following:

$$\textit{slide} = \gamma.z.\sin\beta.\cos\beta \quad (21.2)$$

$$\textit{grip} = \gamma.z.\cos^2\beta \quad (21.3)$$

where slide = shear stress
grip = normal stress
γ = bulk density of soil
z = vertical thickness of soil above potential shear surface derived from the depth normal to the slope and the angle of the slope (hence the $\cos\beta$ in each equation)
β = slope angle

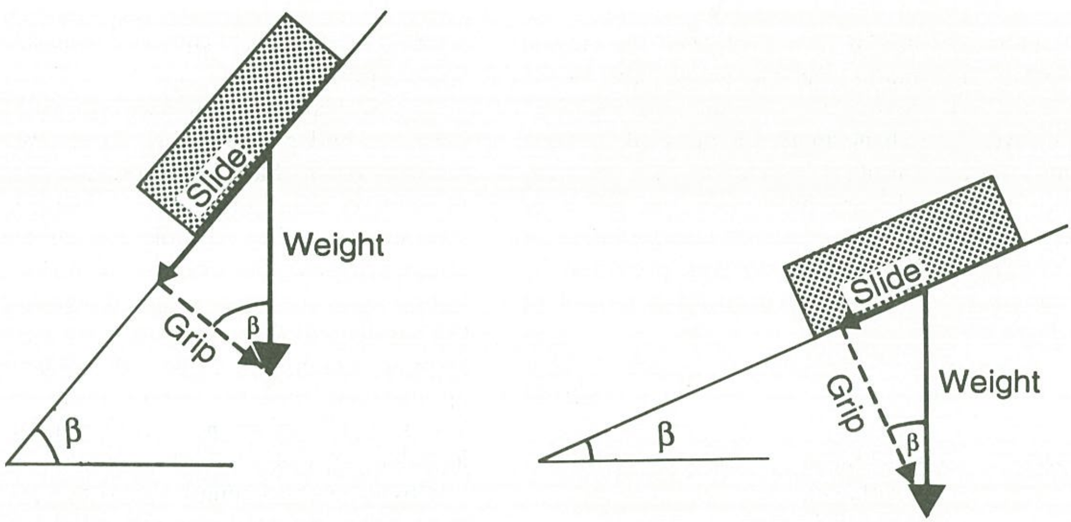

Figure 20.2 The weight of the material on a slope can be resolved into two components: one that promotes sliding and one that "grips" the slope. The ratio of these two stresses is largely a function of the slope angle.

Water in Slopes

Slope materials do not consist solely of solid particles. The regolith also consists of voids (spaces), some of which will contain water. The nature of this pore-water is critical in determining the behaviour of the regolith and its strength. It is therefore important to know something about moisture conditions found in regolith.

In regolith that is not saturated any moisture present is held within the pore spaces at pressures less than atmospheric i.e., it is held by capillary suction or tension (Figure 21.3). The strength of this suction depends on the distribution of pore sizes. Small pores hold water at much greater suctions than large pores, for the same reason that water will rise higher up a thin straw placed vertically in a dish of water than a thick straw. This suction can actually increase the strength of the regolith by effectively "sticking" the particles together.

At some position beneath the surface the pores are completely saturated and the water is held at pressures equal to atmospheric (Figure 21.3). In unconfined aquifers this is usually called the water table or piezometric surface. It commonly follows the general ground surface, only in a more subdued manner. The reason for this is very simple.

Water in the soil has energy from two main sources. Pore-water pressure, the pressure measured by a piezometer, and which in simple systems is proportional to how high water will rise up a standpipe or well drilled into the ground. Water also has additional energy because of its elevation above some arbitrary datum i.e., water higher up the slope will have more energy. Because of the disequilibrium of total water pressure (pore-water pressure + elevation) within a slope, water will flow from areas of high energy to areas of lower energy. In simple systems much of the energy of the groundwater comes from the elevation component of head; this controls groundwater flow, and therefore the piezometric (groundwater) profile tends to mimic surface topography.

Saturated conditions, which are important in slope stability, can exist in three forms. If the

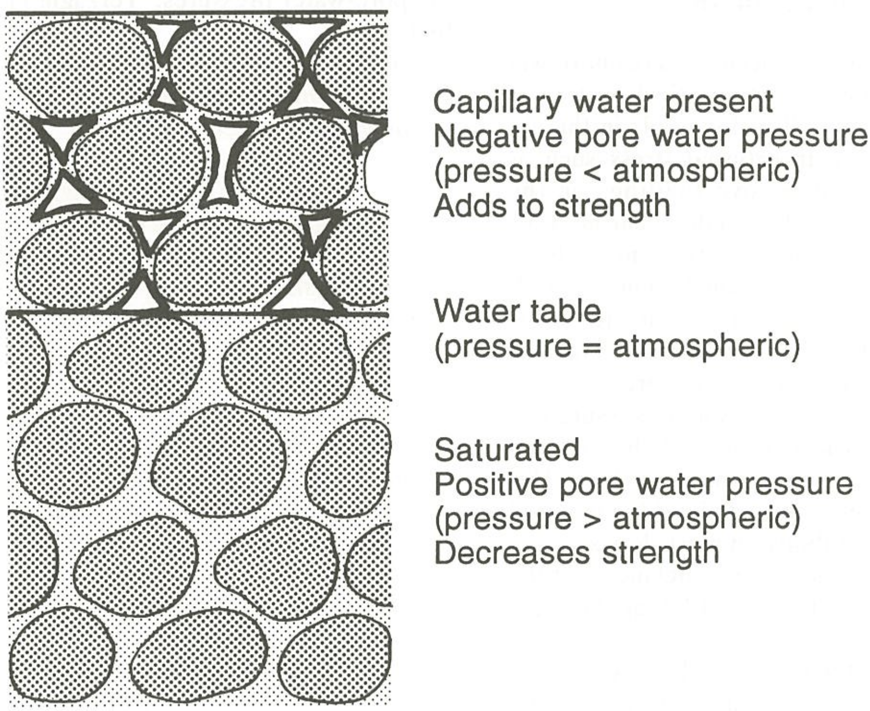

Figure 20.3 Soils consist of solid particles and voids. Depending on the season and availability, these voids contain varying amounts of water which can either increase or decrease the effective strength of the material.

saturated zone is in direct contact with the atmosphere through the pore spaces, then the system is said to be unconfined and the position of the water table will fluctuate with the availability of water. A saturated layer may, however, be confined, trapped between one or more low-permeability layers (aquitards) which restrict pressure equalisation. If this water is tapped, it will rise above the level at which it is found to a height proportional to the pore pressures. In certain situations a saturated layer may be found at higher elevations than the "regional groundwater". This may happen if the saturated layer is isolated from the regional groundwater by a less permeable layer, or during intense rainstorms, where permeable regolith overlies a less permeable bedrock. During a storm precipitation infiltrates rapidly through the regolith, but if it cannot infiltrate the bedrock at the same rate, it "backs up" and saturates the overlying material. Soils saturated in this way normally drain very rapidly once the storm has passed.

Effect of Water on Strength

Since many slope failures occur as a result of wet conditions, water must exert a critical control on stability. For example, if a slope fails in the absence of an increase in external stress such as caused by earthquakes, overloading, or the removal of lateral support, then the shearing resistance (strength) of the material must have decreased i.e., if the "slide" component is not altered and yet failure occurred, then the "grip" component must have been reduced. The two most common ways for this to occur are:

1 An increase in the pore-water pressure i.e., an increase in the buoyancy of the material through a rise in the piezometric surface. This effectively reduces the weight of the material that maintains stability on most slopes.
2 A decrease in cohesion or "stickiness" of the material forming the contacts along the shear plane.

For many years it was argued that water acted as a lubricant, although this has been shown to be generally untrue. For most of the common minerals (quartz, feldspar, calcite) wet surfaces are harder to mobilise than dry ones. Water acts as a lubricant for layer minerals such as biotite, chlorite and montmorillonite, but it is unlikely that a slope, or even the potential shear plane, will consist entirely of these minerals, and the other particles would counteract this effect.

Only a thin film of water is needed for lubrication, and any increase in available water has no additional lubricating effect. This is for the same reason that car engines do not function any better if the engine is "over full" with oil. In humid areas such as New Zealand every sediment usually contains more than enough water for lubrication. Therefore even if water was lubricating the shear surface it would have affected the slope long before it became unstable, and could not be regarded as the trigger of the failure.

Effect of Pore-water Pressures

The most common way in which water reduces the shearing resistance of material is through a rise in the pore-water pressures. Terzaghi (1950) stated that:

Throughout a saturated mass of jointed rock, soil, or sediment, the water which occupies the voids is under pressure. Let:

p = *pressure per unit area at a given point P on a potential surface of sliding, resulting from the weight of solids and water located above the surface, (this is now denoted by σ)*
h = *the piezometric head at that point*
w = *the unit weight of water, and*
ϕ = *the angle of internal friction for the surface of sliding.*

Regarding the relation between these four quantities, soil mechanics has led to the following conclusions. If the potential surface of sliding is located in a layer of sand or silt (which therefore is cohesionless), the shearing resistance s per unit of area at the observation point is equal to:

$$s = (p - hw)\tan\phi$$

Hence, if the piezometric surface rises, h increases, and the shearing resistance (strength) s *decreases. It can even become zero. The action of the water pressure hw can be compared to that of a hydraulic jack. The greater hw, the greater is the part of the total weight of the overburden which is carried by the water, and as soon as hw becomes equal to p the overburden 'floats'. If the material has cohesion, c per unit area, its shearing resistance is equal to the sum of s and the cohesion value c, whence:*

$s = c + (p\text{-}hw)\tan\phi$

The important factor in determining the resistance of regolith to stress acting down-slope is not the total stress induced by its weight, but the difference between the total stress and the pore-water pressures, i.e, the effective stress.

The total stress exerted on a horizontal area within a soil is made up of the pressure transmitted through the chains of particles, and the pressure transmitted through the pore-water. While the total stress is transmitted through both, only the interparticle contacts can resist deformation and failure (Figure 21.4). This is perhaps why large masses, such as ships, can float on water, but as water lacks shear strength it will flow away when poured from a container. Pore-water can support vertical pressure but has no shear strength, so if the shear plane is below the water table the "grip" component of strength is reduced, leading to failure.

Beneath the water table any object (e.g., the overlying soil material) is subjected to a buoyancy or relief of weight equal to the weight of water displaced. Equations 21.2 and 21.3 must be modified to take account of the stress carried on the pore fluid. Therefore our equations for the two stresses acting on the material become:

$$\text{Shear stress } (\tau) = \gamma.z.\sin\beta.\cos\beta \qquad (21.4)$$

$$\text{Effective normal stress } (\sigma') = (\gamma\text{-}M\gamma_w).z.\cos^2\beta \qquad (21.5)$$

where M = ratio of the water table height to the height to the ground surface above the failure plane

γ_w = bulk density of water

All other variables are as in equations 21.2 and 21.3 When parameters take account of pore-water pressures, that is they are effective strength parameters, they have a ' superscript.

As the water table rises towards the surface, M gets closer to unity and therefore the effective weight of the overlying material decreases (Figure 21.5).

If we add to equation 21.5 the parameters for the cohesion and coefficient of friction of the material, we can calculate the relative stability of slopes (F) subject to infinite planar slides by looking at the ratio of shear strength to shear stress:

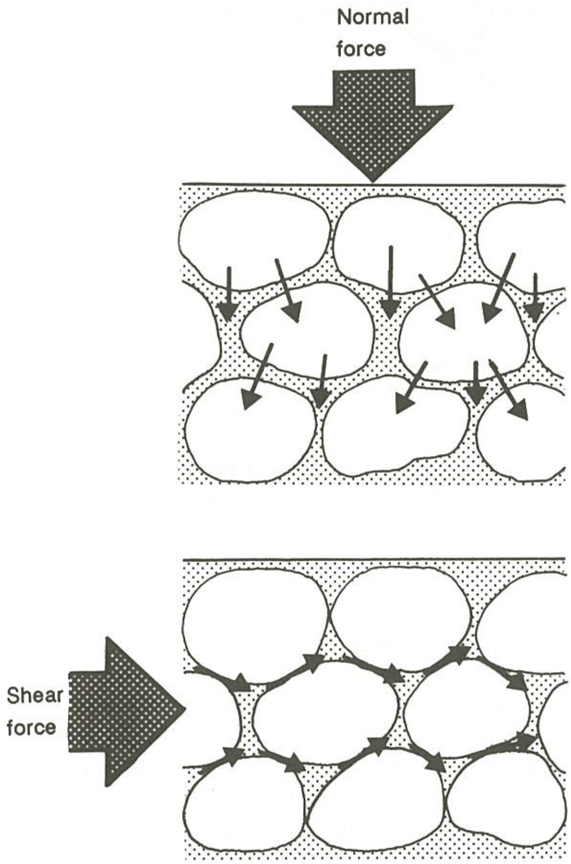

Figure 20.4 Both water and the solid contacts between particles can carry any normal stress applied to the material, but only the solid contacts can withstand any shear stress. Water in a saturated sample thus reduces the strength of the material.

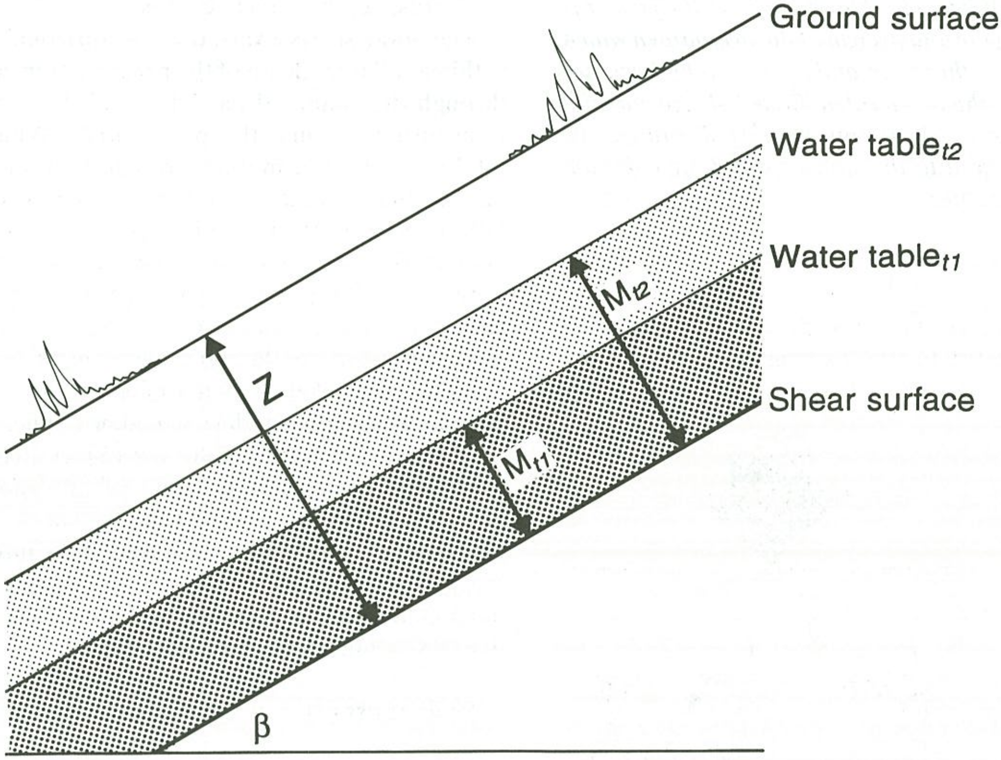

Figure 20.5 The position of the water table influences the effective stresses and therefore the strength of a slope. As the pore-water pressures increase, the frictional strength of the material is reduced.

$$F = \frac{c' + (\gamma - M\gamma_w)\, z.\cos^2\beta.\tan\varphi}{\gamma z.\sin\beta.\cos\beta} \qquad (21.6)$$

From this equation, and others like it for different types of slope failure, it is possible to analyse how a change in one variable will affect the stability of the slope. Usually the only variable that is likely to change significantly within a short period of time is the pore-water pressure. Although changes in water content have their most dramatic effect on the effective normal stress, and hence the frictional strength of the material, they also affect the cohesion and bulk density.

Other Effects of Water

Moist soils often contain additional cohesion because the water in the pores is held under tension. Should moisture levels rise, either during rain or a wet winter, an increasing number of pores become saturated, reducing the zone of negative pressures and therefore the strength.

The effect of increasing water content on electrochemical cohesion is harder to visualise. Because of the electrochemical charges on the surfaces of clay particles, water molecules which are bipolar (they have a negative charge at one end and a positive change at the other) orient themselves between the layers of clay particles (Figure 21.6). As the amount of water increases, the "buffer" of water between the clay layers increases, with the result that the electrochemical attraction between the clay layers is reduced. Depending on the strength of the electrochemical charges and the availability of water, the clay particles may be bound tightly together or they may be bound loosely in a matrix of water. In this latter condition, because the bonds are not rigid, the clay particles can move under very low shear stresses. In some soil types, such as those forming much of New Zealand's pastoral hill country, just changing the water content can significantly change the strength of the material.

Joints, fractures, or other structural voids in a rock mass forming a slope can act in the same way as the pores in a regolith. Water in these joints ("cleft water") has the same effect as pore-water in porous regoliths, and affects the stability equation in a similar manner. Cleft water exerts pressure on the sides of the joint and reduces the effective weight, the friction, and therefore the strength of the overlying rock mass. Depending on the orientation of the joint planes the cleft water may also "push" against the front of the rock slope, further decreasing the strength and increasing the possibility of failure. The effects of cleft water are shown in Figure 21.7.

Summary

According to Crozier (1986) changes in water content can quickly and dramatically affect the stability of slopes, triggering, re-initiating, and accelerating more landslides than any other factor. An increase in water content decreases stability in one or more of the following ways:

1 Increasing pore-water pressures within the voids, reducing the frictional component of strength.
2 Decreasing cohesion and suction, reducing the cohesive strength of the material.
3 Increasing cleft water pressure within joints, voids and fissures, reducing the frictional

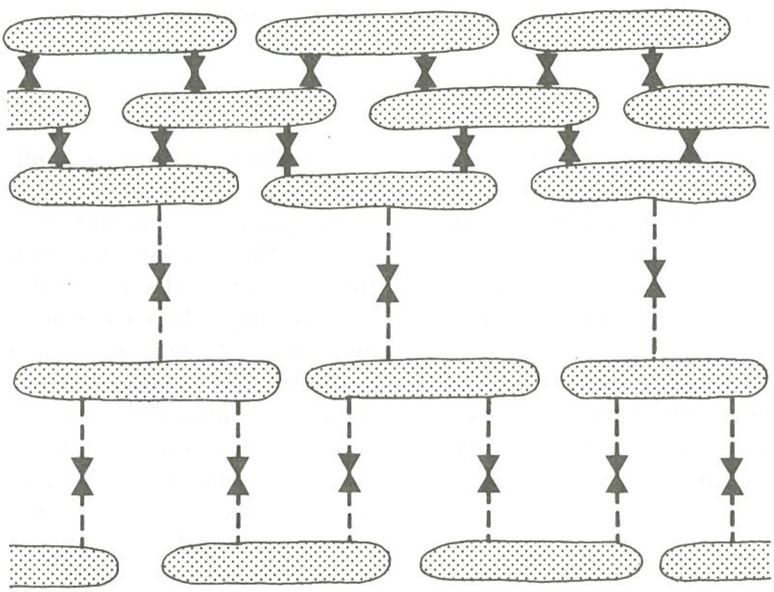

Figure 20.6 The strength of attraction between cohesive particles can vary with the water content of the material. Additional water can increase particle separation, leading to a reduction in strength.

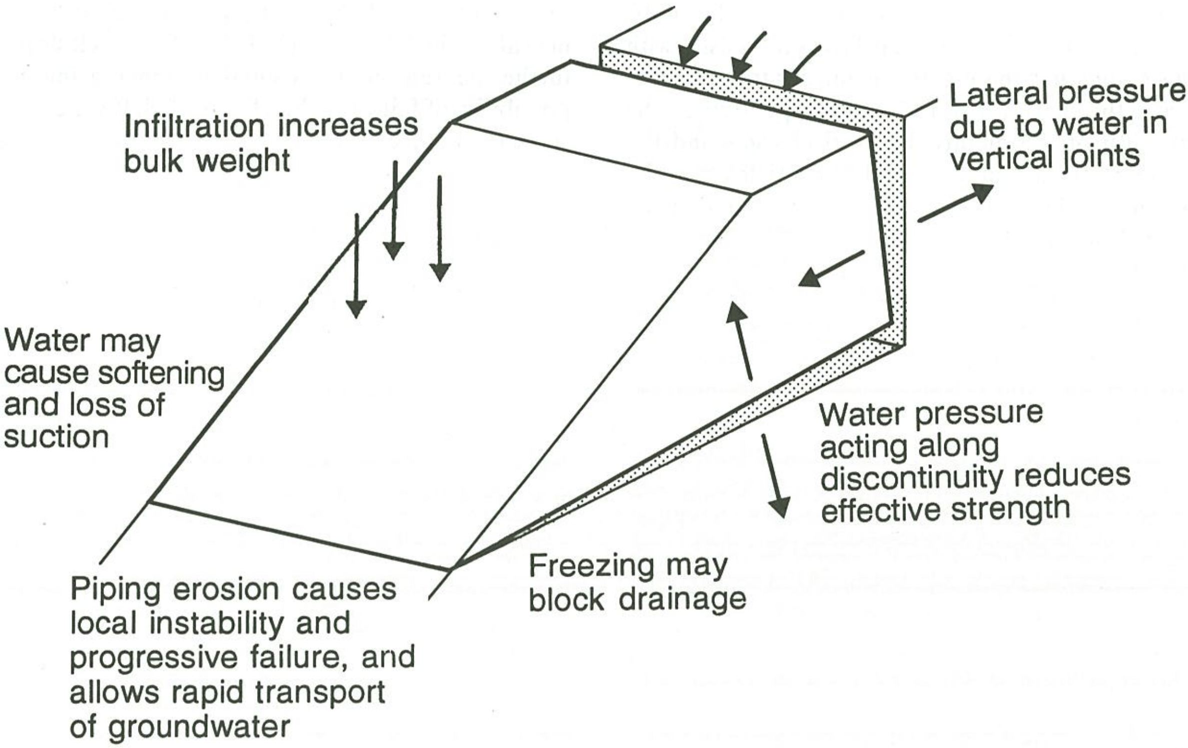

Figure 20.7 Cleft water in the joints of rock can reduce its strength in a number of ways (After Hencher, 1987).

strength along these planes of weakness.

4 Generating seepage pressures which set up a drag stress in the direction of water percolation, contributing to the shearing stress.

5 Lubricating certain types of minerals, reducing frictional strength across the mineral surfaces.

6 Increasing the weight of the material. This can initiate failure only in cohesive slope material or in fractured material with stability maintained solely by root cohesion.

Natural Materials

Some materials forming natural slopes exhibit different responses to shear stress, depending on the rate at which the stress is applied and the normal stress pushing the particles together.

Many mass-movement features in New Zealand occur in Pliocene mudstone that consists of a mixture of chemically inert silts and sands, and cohesive clays. This material has both its cohesive and frictional components of strength disrupted, and behaves in either a cohesive or frictional manner depending on the stress conditions.

Under high normal stress, the material's behaviour is controlled largely by friction and thus the influence of the silts and sands dominate. Their frictional strength must be overcome before any movement can take place. If, however, intact material does fail the degree of packing (which effects the number and strength of the grain-to-grain contacts) is reduced, and the thickness of the films of water surrounding the solid particles increases. This same material will then deform (creep) at stress levels very much lower than those required to overcome its initial "frictional" strength. Once failure has occurred, the material changes in such a way that instability will occur at much lower stress levels, making it much harder to stabilise the slope (McConchie, 1986).

Laboratory tests on this commonly occurring mudstone, using both a shear box and triaxial device, showed that material that had already failed yielded under very low shear stresses when tested under low normal stresses (40kPa), because of the thick layer of water surrounding individual particles. Under low normal loads this material would deform as a 'plastic' indefinitely. At greater normal stress levels (80kPa) this same material behaved like a typical undisturbed sample with frictional strength, and a measurable peak strength. All the samples tested that had not been subjected to previous failure, no matter what normal loads where used, showed a mode of failure typical of a frictional material.

For intact rock the particles are in contact and the behaviour of the material is controlled by the frictional component of strength. After initial failure the material has more, or larger, voids and a saturated sample can therefore hold more water. This allows the material to respond under low normal and shear stresses in a manner akin to a cohesive material.

Shear strength is highly correlated with moisture content. A slight increase in moisture content causes a large decrease in shear strength. For example, the undrained shear strength (which takes no account of the effect of the porewater) increases from about 10-15kPa at 45% moisture content to approximately 300kPa at 20% (Figure 21.8). The compressive strength increases from being so low as to be immeasurable (although probably 50kPa at 30% moisture content) to 500kPa at 20%, 2000kPa at 10%, and 4500kPa at 5% (McConchie, 1986).

Therefore, while the water content is important in looking at the behaviour of material, particularly when looking at short term changes in stability, the stress levels and history of material should not be overlooked. Laboratory tests to determine strength properties of real slopes must be carried out in a manner that accurately replicates the real stress levels and history of shearing.

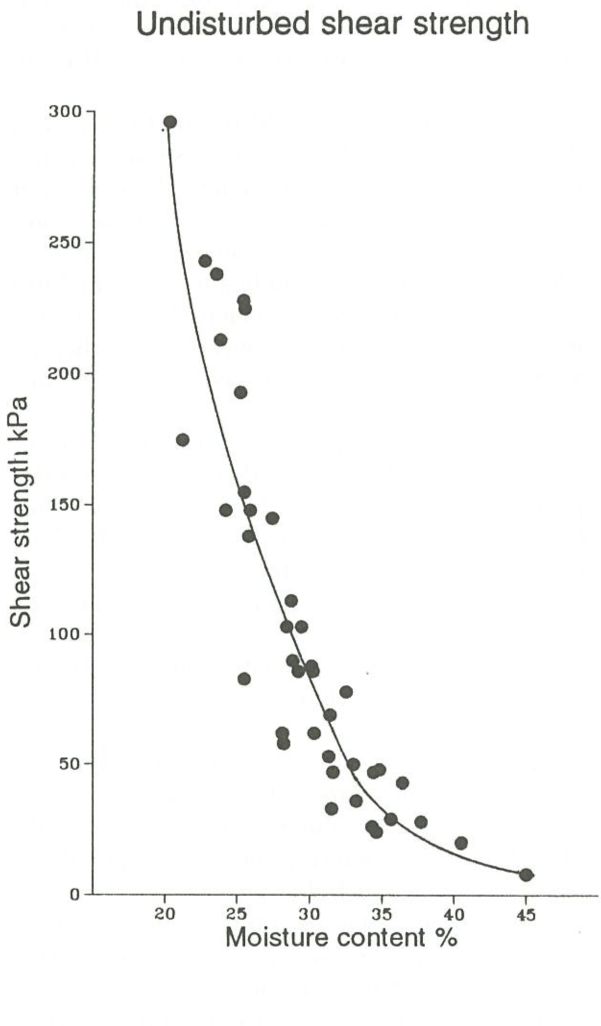

Figure 20.8 In some soils, such as this example from the eastern Wairarapa, increasing the moisture content can decrease the strength markedly. A change in moisture content of only 20% reduces the strength by more than 90%.

Groundwater flow

Groundwater conditions are among the prime determinants of the effective normal stress acting within a regolith and therefore its strength. While it is relatively easy to establish models of groundwater flow in 'artificial' or 'design' slopes, the irregularities in natural slopes restrict the usefulness of theoretical modelling in many field situations. In the study of landslides, the modelling of groundwater conditions is complex, as the topography of slump features complicates the groundwater circulation system.

In a study of an earthflow in the eastern Wairarapa, McConchie (1986) has shown that, not only is water critical in determining the stability of the slope, but that once failure has occurred, changes in the slope promote continuing instability. The major change is a modification of slope form from planar to concave, leading to the mass movement feature acting as its own catchment, thereby altering soil moisture and groundwater conditions within the slope.

Many studies have shown a relationship between total rainfall and its distribution, and the position and fluctuations of the water table. McConchie (1986) measured the depth to the water table recorded in 60 piezometers installed across the earthflow, generating an 'average' trend for the slope (Figure 21.9). He found that the earthflow exerts a major control on groundwater in the slope, and that it appears to have a saturated capacity above which the water table does not rise, at least not during the years for which data are available. This upper limit of the height of the water table is reached very rapidly with the onset of winter, and is maintained for several months until slow drainage during the summer begins. The response of a particular monitoring site is largely controlled by the relationship of that site to slope morphology and any hydraulic controls (such as aquitards) within the unstable material. Because the earthflow acts like a large sponge, sites at the top of the slope and surrounding the feature are more variable in their groundwater conditions than sites within the earthflow material.

Flow Patterns

Many landslides control the groundwater flow by concentrating and channelling soil moisture and groundwater towards the centre and toe of the feature. This can be illustrated by looking at the variation in total head or energy of the groundwater in the slope.

A traditional approach is to depict pressure gradients with flow nets. These are usually drawn for 'design' slopes in which not only are the flow characteristics known but conditions are assumed to be relatively homogeneous. However, landslides

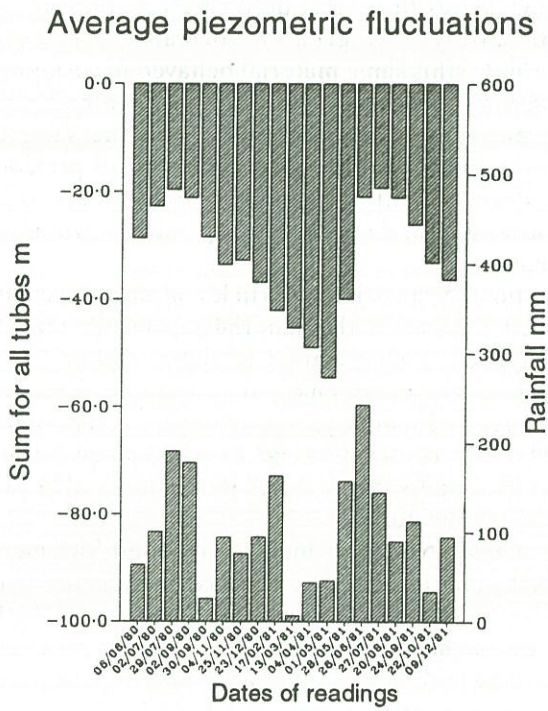

Figure 20.9 The position of the water table is controlled by the effective precipitation i.e., precipitation minus evapotranspiration. The slow drainage of groundwater during summer and rapid recharge with the onset of winter in a slope undergoing failure were highlighted by measuring the depth to the water table at 60 locations within an earthflow in the eastern Wairarapa (McConchie, 1986).

and natural slopes are far from homogeneous. Groundwater conditions can be accurately modelled only if the total head is known at a large number of points in the slope. In a study of an earthflow where the total head was known for 60 positions within the slope it was possible to construct a three-dimensional model showing groundwater flow (McConchie, 1986). Because several piezometers were installed vertically in each bore hole it was possible to measure the "unit vertical pressure difference" between the piezometers which indicates the direction and relative magnitude of the vertical component of flow.

Figure 21.10 shows flow conditions when the water table is highest (winter) and when it is lowest (summer). From these two plots, several characteristics of groundwater flow are apparent. First, groundwater flow occurs at right angles to the piezometric isolines and from areas of high pressure towards those of lower pressure. Second, the isolines conform to the morphology of the slope and so mimic the topographic contours. The pressure gradient is steepest on the steeper terrain and levels out both at the drainage divide and at the base of the slope, where the stream acts as a local base-level for groundwater flow. This pattern reflects the control the elevation component of head exerts on the total head pattern, at least in simple groundwater systems.

When the water table is highest, the vertical component of flow is generally into the ground (as one would expect) except at three locations. One area of apparent upward flow is in the depression below the crown scarp where ponding is evident from winter until mid summer. The other two sites with an upward component of flow are at the base of the slope. In late summer the pressure distribution shows two major changes. The first is a decrease in the pressure gradient and the second is a more active channelling of groundwater into the earthflow.

Once the features have formed, the morphology of an earthflow, and probably other types of landslides, controls the drainage patterns within the slope and encourages the maintenance of high water tables and pore-water pressures, particularly at the toe where they have the greatest affect on stability.

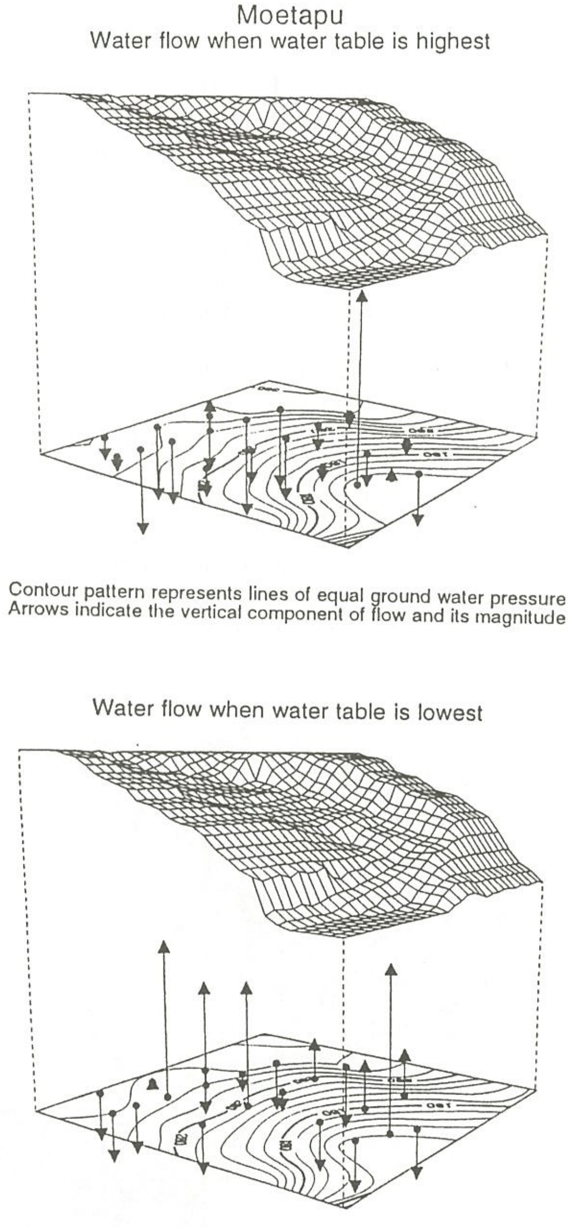

Figure 20.10 Groundwater flow patterns for an earthflow in eastern Wairarapa. Three-dimensional models of groundwater pressure allow the interpretation of flow patterns. Because of natural variability the development of such models for landslides is difficult without a large number of observations.

Rural landslipping

Fast Shallow Failures

During the winter of 1977 the hill country and plains of the Wairarapa experienced unusually wet conditions. Between June and September there was over 700 mm of rain, approximately 150% of the normal rainfall for this period. This excess rainfall was distributed unevenly, with September having 325% more rain than usual near the western edge of the hill country. A consequence of this prolonged wet period was widespread and severe shallow soil slippage in the hill country east of the plain (Figure 21.11).

The loss in annual income for farmers from the 1400 km2 most severely affected was estimated to be approximately $1M in the first year. The impact of this event, and others like it in the past, however, will persist as the loss of soil from the slopes reduces potential productivity from grazing. It could be argued that the soils of these slopes are being "mined" to maintain a pastoral regime.

The slope failures during this episode involved the initial failure of the regolith followed by failure of the restraining turf mat. The regolith was 'saturated' and behaved as a "fluid". This sequence of failure was described by witnesses and is supported by the presence of incipient slips consisting of an ovate form with an unruptured turf bulge at the toe and a number of crescentic tension cracks at the head (Owen, 1981 a & b).

Because of the severity of landslipping in this episode an intensive study was undertaken in one 20 ha catchment to identify the cause of such widespread slope failures. The Pakaraka catchment was one of the most severely affected catchments in the area with 4.78 slips/ha, when the average for the general area was 0.98 slips/ha (Crozier *et al.* 1980).

This study (Owen, 1981 a & b) showed that the sunny northerly aspects were much more susceptible to failure than the shady slopes. The slope failures were most common in regolith undis-

Figure 20.11 Shallow regolith failures, such as these on the east coast of the North Island, are all too common in New Zealand's rural landscape. Intense rainstorms or prolonged wet periods leading to such failures occur almost every year.

turbed prior to 1977, that is the landslips occurred in "new" locations.

When compared to other New Zealand soils, the soils of the sunny aspects have particularly low liquid limits (the moisture content at which a remoulded sample changes from behaving as a plastic to a liquid) and low plastic limits (the boundary between the plastic and semi-solid states). Consequently when remoulded the Pakaraka soils lose strength rapidly with rising water content, and that loss of strength is particularly rapid once the soil is above its plastic limit. Tests showed that the strength of the undisturbed regolith of the Pakaraka basin also differs with slope aspect. At normal winter moisture content, the soils of shady slopes are a little stronger than those of sunny slopes, and at higher moisture contents this difference is accentuated; soils of the sunny aspect lose strength more readily with rising water content than do the soils of the shady aspect. On sunny slopes the turf mat is generally the most resistant part of the regolith (Owen, 1981 a & b).

In 1977 prolonged rainfall and low evapotranspiration allowed the normally drier sunny slopes to achieve a moisture content close to that of the shady slopes. At these moisture contents they were significantly weaker and therefore failed more readily. The high incidence of flowage of these slips was because the sunny slopes have a water content above their liquid limit when saturated. Once failure has occurred and the turf mat ruptures, the "fluid" material inside can flow easily down the relatively steep slopes (24-40^{o}) and out onto the valley floors. The shady slopes when saturated are still below their liquid limit. These sunny slopes had not failed previously because the moisture contents of their soils were kept low by the prevailing northwesterly winds and prolonged sunshine, both absent during the winter of 1977.

Slow Deep-seated Failures

Whereas most landslides start moving suddenly and usually come to a rest in a matter of minutes, or at least within a few hours, some forms of mass movement creep continuously throughout the winter and start of spring when moisture levels within the slope are high. Many of these features are classed as earthflows. Over one million hectares, or 4% of New Zealand's productive land is undergoing some form of active slope failure through earthflow (Eyles, 1983).

There are extensive earthflows on the soft, gently sloping clays on the east coast of the North Island. The slopes have a characteristic hummocky appearance, and small ponds occupy the depressions for much of the year. In describing the movement of such features Campbell (1951) wrote:

Blocks of turf seem to float down on the viscous underlying mass and stand out individually with wide, knee-deep, cracks between them while many are partially overturned. They occur after heavy rains and many involve the complete drainage depression running down the slope and it appears that the mass of material was super-saturated and the resistance of the turf mat finally gave way. Gradually the flows dry out and remain stable until the next rains.

From the intensive study of one such earthflow (Figure 21.12) in the eastern Wairarapa it would appear that failure occurs through two separate, though interacting, mechanisms, both of which are controlled by the amount of water in the slope. One involves the mobilisation of the predominantly frictional strength of the material. The other mobilises the cohesive strength which interacts with the compressive strength of the material and the hydrostatic pressures at the toe (McConchie, 1986).

In this example, the initial failure of the intact mudstone appears to have been caused by the downcutting of the stream at the base of the slope. This increased the shear stresses acting in the slope until the entire shear strength of the intact material was mobilised and failure occurred. This failure process can be modelled using limit equilibrium analysis (equation 21.6) adjusted to take account of the shape of the failure plane and variations in properties within the slope. For this particular slope geometry and material strength, the water table would have needed to be at the ground surface for failure to occur (McConchie, 1986).

Because the slope failed under low effective stresses, a result of the high pore-water pressures

Figure 20.12 A deep-seated seasonal earthflow in the eastern Wairarapa. This type of slope instability affects over one million hectares in New Zealand.

existing at the time of failure, the unstable material's properties were modified. Its void ratio increased and its bulk density decreased, reducing interparticle contacts, and therefore the frictional strength of the material. The thickness of the layers of water surrounding the silt and clay particles also increased. These changes allow the particles, if not constrained in some manner, to flow under very low stresses, such as those imposed simply by gravity on low-angled slopes. During summer this plastic material remains stable because it is confined within a "high strength" envelope formed by the dry soil at the surface of the toe. Even at this time, though, if a hole is drilled through the crust, the material inside is sufficiently plastic to flow into and close the hole within a few minutes.

During wetter conditions the higher soil moisture contents reduce the compressive strength of the exposed material at the toe (by up to three orders of magnitude) while the higher water table increases the stresses acting on the face. When the stresses acting on the face are greater than its compressive strength, the material begins to creep forward. The rate of creep is governed largely by soil moisture and groundwater conditions within the soil. During winter when pore-water pressures are greatest (thus reducing the normal stress and increasing hydrostatic pressures) and soil moisture is highest, this creep can affect the whole earthflow.

When sufficient material is removed from the toe, the slope behind loses its lateral support and intact material further up the slope fails by the same mechanism as the initial failure. In this manner the earthflow develops retrogressively upslope.

Urban Landslipping

Shallow Regolith Slides

Recurrent episodes of landslipping have occurred within Wellington City this century, related to increasing annual rainfalls and suburban expansion onto steep slopes. The recurrence interval of such periods of slipping is as low as 3.4 years (Eyles, 1979). These landslides reflect disequilibrium in the geomorphic system brought about by climate change since the last glacial 20,000 years ago and human modifications of the landscape. The landslides commonly occur in colluvium-filled bedrock depressions (fossil gullies) rather than on ridges, spurs, or sites with planar bedrock surfaces (Crozier *et al.,* 1990). Regolith at these sites generally exceeds the critical depth required for instability and is therefore more susceptible to landsliding. The increased occurrence of failure of these sites in recent time appears to have been brought about by deforestation removing root cohesion. Tectonic uplift, resulting in steeper slopes, has also changed the stress conditions within the slopes and promoted instability. While these are the preparatory factors of instability the actual trigger of these landslides usually involves water.

The majority of landslips in Wellington are surficial, weathering-controlled, debris slides involving shallow regolith and intensely jointed, weathered greywacke (Eyles *et al.,* 1978; McConchie, 1980). The slopes most susceptible to failure have been 'cut' for construction purposes, often receive localised drainage, and are predominantly north-facing.

While the magnitude of the response of the environment to rain has increased because of human activities, it does not mean that intense landslipping did not occur under natural conditions. The December 1976 storm demonstrated that natural slopes can, in extreme conditions, experience landslips. During this event, Stokes Valley, a 12 km^2 catchment in Lower Hutt, experienced 78 landslips and severe flooding as a result of landslip debris blocking the drainage system (Figure 21.13). Those sites most prone to failure were high in the drainage basins, on slopes greater than 21^{o}, where drainage was convergent, and the regolith was composed of greywacke-derived colluvium overlying the bedrock (McConchie, 1980).

Periods of slipping in Wellington appear to result from either very wet winters (e.g., 1974) or intense rain (e.g., 20 December 1976), so water acts as the trigger in many of these landslides. This is also indicated by the number of landslides in locations prone to rapid build up of pore-water pressures within the regolith or highly jointed bedrock.

Figure 20.13 An intense rainstorm on the 20 December 1976 led to widespread landslipping in Wellington. While only one life was lost numerous houses suffered damage such as these in StokesValley. (Photograph: Courtesy of The Evening Post)

On many sites in Wellington a relatively permeable regolith overlies impermeable bedrock. Perched water tables develop during heavy rain leading to a reduction in normal strength and consequently shear strength. Sites prone to extreme fluctuations in pore-water pressures (where the water table can even reach the surface) are those located in concave depressions or those with convergent drainage. Cleft water pressures within the bedrock also can reduce strength.

There appear to be three threshold values of rainfall above which serious slipping occurs. On "cut-and-fill" slopes modified for urban development slipping occurs with either a four-month rainfall of between 750 and 800 mm (susceptibility increasing towards the end of the wet period) or, following relatively dry antecedent conditions, a 24-hour rain storm above 120 mm. Natural slopes under grass, scrub, or forest appear to be more stable, and a 24-hour rainfall of from 200 to 250 mm is required to initiate slipping (Eyles *et al.,* 1978).

The accurate prediction of landslides is particularly important in urban areas where property and safety are at stake. However, because of the highly variable nature of the regolith, bedrock, and precipitation patterns (therefore groundwater responses) in Wellington, effort has focused on developing generalised indices of instability. One of the more successful was developed by Crozier and Eyles (1980). Because the size of a slip-triggering storm is related to the pre-storm moisture conditions, their 'Antecedent Excess Rainfall' model integrates both antecedent moisture conditions and rainfall parameters. The model assumes that the soil can store the equivalent of 120 mm of rainfall and that any gravitational water can drain from the regolith within 10 days. For a particular day there will be an excess or deficit of soil moisture, depending on climatic conditions in the preceding 10 days. Using this simple model and daily rainfall, a clear separation was found between days with slips and those without (Figure 21.14) for Wellington City in 1974. The model therefore has considerable predictive capability.

This simple model, while not explaining failure processes, does integrate a range of properties and shows the role of water in triggering mass movements in Wellington City.

Deep-seated Failures

Perhaps the most dramatic landslide in New Zealand's recorded history was at Abbotsford, Dunedin. On 8 August 1979, 18 hectares of hillside slid downhill, destroying 69 homes and affecting hundreds of people (Figure 21.15).

While the final collapse of the slope was dramatic, its potential instability had been discussed for over 20 years, and the initial movement of the slope was detected ten years earlier. Movement of the slope began to accelerate in early May 1979; a water-supply pipe was pulled apart in Mitchell St and nearby houses suffered damage. Throughout July there was growing evidence that a very large block of land was moving and that the movement was accelerating. At about 9 pm on 8 August, a block, with an estimated volume of 4.8 to 5.4 million cubic metres, began to move rapidly in a south-easterly direction at a rate of about one metre per minute. The main movement was over in less than an hour, but in that time the block had slid almost 50 metres (Commission of Inquiry, 1980).

This landslide was distinctive in New Zealand in that the material that slid was relatively stable, but it overlay a weaker layer dipping gently downslope.

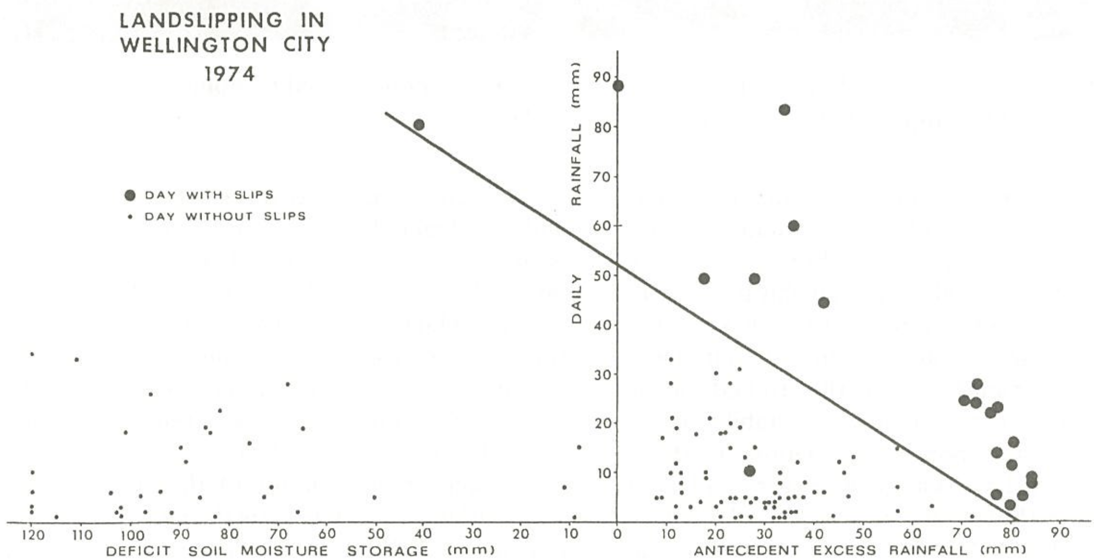

Figure 20.14 Crozier and Eyles (1980) model has been used to predict periods of slope instability. Periods of landslipping are correlated with the moisture content of slope deposits. Greater daily rainfalls are required to trigger slope failures under dry antecedent conditions than if the soils are already wet.

Figure 20.15 The Abbotsford landslide in Dunedin in August 1979 destroyed 69 homes, affecting over 200 persons. (Photograph: Courtesy of The Otago Daily Times)

The Commission of Inquiry found that the failure plane occurred just below the boundary of the Green Island and Abbotsford Formations. This zone contains many thin beds of highly plastic, silty clay. Careful examination showed "slickensides" (smooth polished surfaces usually associated with shearing movement). It was the strength of this very thin layer that controlled the stability of the entire slope in response to variations in stress. This structural weakness, however, had been present for thousands of years, and while it predisposed the slope to failure it was not the actual trigger.

While the slope failure coincided with a period of wetter than normal conditions, the first indications of movement appeared in the 1960s at the end of a particularly dry period. There was therefore considerable debate as to the exact cause of failure, and this highlighted the difficulty of separating what Crozier (1986) has called the preparatory factors from the actual trigger of instability.

Some blamed the removal of lateral support (buttressing) from the toe, caused by the excavation and quarrying of material from Harrison's Pit, for the instability. Analyses after the failure showed that the removal of this material would have changed the stability of the slope by only 1-2%. While this is only a small change, it is worth remembering that even a 0.1% change, if it tips the balance, could be the trigger of instability.

Not discounting the preparatory factors, the role of water in the final failure was still critical. First, the clay layers that formed the failure plane contained high percentages of montmorillonite.

This is an expanding-lattice clay, and materials with a high proportion of this clay have high plasticity. When wet they have low shear strength and so are prone to sliding and other forms of mass movement. The solid particles in materials of this nature are surrounded by films of water which inhibit solid-solid contacts and reduce the overall strength of the material by allowing the "solid" particles to deform within semi-rigid water films. Montmorillonite is also one of the few minerals (the layer-lattice minerals) where water acts as a lubricant. At the depth of the shear plane, however, the effect of water on the strength may have been constant, since the material was probably continuously saturated. However, slow creep over time can lead to a reduction in the overall stability of a material.

The main effect of water that is likely to vary through time, and therefore be the final trigger, is the groundwater and pore-water pressures operating on the shear plane. It was recognised by the Commission of Inquiry that groundwater levels were probably high when the landslide occurred because of higher-than-average rainfall over the preceding 10 years. However, it was also argued that these conditions would have been equalled, and probably exceeded, in the past without causing failure. As groundwater levels rise, more normal stress is carried on the pore-water, reducing the effective stress and therefore the strength. It was shown that a 1-m rise in the water table in this slope would have reduced the stability by about 3%. It was also argued that leakage from Dunedin City's Water Main might have augmented the unusually high groundwater conditions. Although data are sparse it was suggested that leakage from the main could have been equivalent to about one third of the annual infiltration of rainfall. Analyses showed that this could have reduced the stability of the slope by a further 1.5%.

Once movement is initiated, fissures in the surface provide ready access for rainfall to reach the groundwater. Thus, once movement starts, conditions encourage higher groundwater levels, compounding the instability. While the role of water in this failure is not clearcut, and water was not the only factor leading to instability, it was probably the final trigger.

Stabilisation of Landslides

The Clyde Power Project on the Clutha River includes the largest and most ambitious slope stabilisation project ever undertaken. Of a total project cost of \$1.4 billion, approximately \$0.4 billion has been spent on the investigation and stabilisation of a series of ancient landslides in the Cromwell Gorge.

Approximately 25% of the shoreline of Lake Dunstan which will form behind the Clyde Dam will be bordered by extensive deep-seated "creeping" landslides formed mainly in schist bedrock, schist debris, and colluvium (Figure 21.16). The slides are massive, having a relief of 300 to 500 m (often involving the entire valley side), a thickness of up to 200 m, and slope lengths of up to several kilometers (Table 21.1). Currently the landslides are either dormant (inactive) or are creeping only at very slow rates (< 20 mm.y^{-1}), and they do not appear to have undergone any large-scale rapid movements.

While these landslides are ancient features (being formed at least 50,000-150,000 years ago) their movement has been shown to accelerate in response to heavy rainfall and higher groundwater conditions. This has led to concern over the response and potential risk of these landslides once Lake Dunstan has been filled, especially since some to the toes of the landslides will be submerged by up to 62 m (Gillon and Hancox, 1991).

Initially very little was known of the groundwater system in the area, and it was assumed to be reasonably simple, rising gently away from the Clutha River. Drilling investigations, however, have shown that a single landslide can contain both confined and unconfined aquifers, aquitards, near-surface perched water tables, and high-pressure groundwater systems. These form in response to variations in the unstable material and the numerous faults and shear zones formed by previous movement. To assist in defining the groundwater pattern 992 piezometers have been installed in the Cromwell Gorge landslides (Macfarlane *et al.*, 1991).

The effects of lake filling on the stability of existing slides are well established. Inundation of jointed rock slides with planar failure surfaces causes a decrease in stability. This decrease

Slide Name	Physical Characteristics							Groundwater Conditions	Slide Movements		Summary Of Main Remedial Measures
	Area (ha)	Volume (10^6 m^3)	Slope Length (m)	Toe Width (m)	Max. Thick-ness (m)	Average Slope Angle (^{o})	Depth of Toe Sub (m)		Average (mm/yr)	Max. (mm/yr)	
Clyde	120	50	1050	800	70	22-28	25	Normal, dry	I/NC	-	Toe buttress (dam spoil); D/drive (1020 m); D/drilling (4000 m)
Jackson Creek	23	5	700	350	46	28-30	35	Both P & C systems	12	-	Toe buttress (1.18M m^3); D/drives (8100 m); D/drilling (8400 m)
Hintons	22	8	770	580	70	34	47	Normal UC s/basal	ND	-	Buttress effect of new SH 8; D/drilling (2500 m)
Two Bridges	21	7	440	530	80	30	49	Normal, small P/WT	ND	-	Buttress effect of new SH 8; D/drive (1120 m); D/drilling (10,300 m) [mainly for Dunlays Slide],
Dunlays	74	55	1200	650	120	23-32	49	P/WTs; C s/basal WT	ND	-	
Flying Fox	14	5	400	400	60	27	47	Slide dry; some C s/b	15	-	Buttress effect of new SH 8; D/drilling (2100 m)
No. 5 Creek	126	60	1700	800	100	19-35	40	C s/basal gw; P in sl	5	20	Buttress effect of new SH 8; D/drive (710 m); D/drilling (4200 m)
Nine Mile Ck Ds*	900	>1000	3-4 km	2700	200	16-27	40	C s/basal gw; P in sl	3	7	D/drives (3280 m); D/drilling (21,900 m); toe buttress (1.38M m^3)
Nine Mile Ck Us	300	240	1700	1200	180	19-25	40	C & P gw int & s/basal	8	16	D/drives (4820 m); D/drilling (39,500 m)
Cairnmuir	100	10	990	1300	83	25	nil	C s/basal; P in sl	30	90	D/drive (680 m); D/drilling (5000 m); surface reveg & drainage sys
Brewery Creek	200	175	1100	2000	140	26	40	UC s/basal; sm P in sl	20 (AP)	-	D/drives (pumped & gravity 3620 m); D/drill (21,900 m); toe cutoff blanket & buttress (1.71M m^3); toe grout curtain (12,880 m)
Cromwell	7.5	3	250	300	35	31	32	Slide dry	10	40	Toe buttress (0.3M m^3)
Miners (rockfall)	5	0.75	400	150	30	40	15	Dry pos P in sl	nil	nil	Remedial measures considered unnecessary
Cornish Point	13	4	550	550	20	30	15	Slide dry	7	-	Remedial measures considered unnecessary

NOTES:

1. Toe width relates to section of slide within the proposed reservoir. Where ranges are given for slope angles, these generally indicate flatter upper slopes, and steeper lower slopes.
2. * Figures given for the Nine Mile Creek Slide Downstream Segment are for the entire slide mass. Investigations and remedial measures, however, relate mainly to the downslope areas of the slide.
3. Abbreviations used are as follows: (C)-confined groundwater system; (UC)-unconfined groundwater; (P)-perched groundwater; (gw)-groundwater; (s/basal)-sub basal (or below slide) gw; (sm)-small; (D/drive)-drainage drive; (D/drill)-drainage drilling; (sl)-slide; (I/A)-inactive; (I/NC)-slide movement indicated but not confirmed; (ND)-no slide movement detected outside error limits; (AP)-movement detected only on frontal portion activated by road construction.
4. Totals given for investigations include all work completed or programmed as at June 1991 (from Gillon, Denton, and Macfarlane, 1992). These totals indicate all work with an investigation input carried out during stabilisation work, as well as during earlier investigations.

(Modified from Gillon & Hancox 1991).

Table 20.1 Summary of characteristics and investigations of major landslides in the Cromwell Gorge

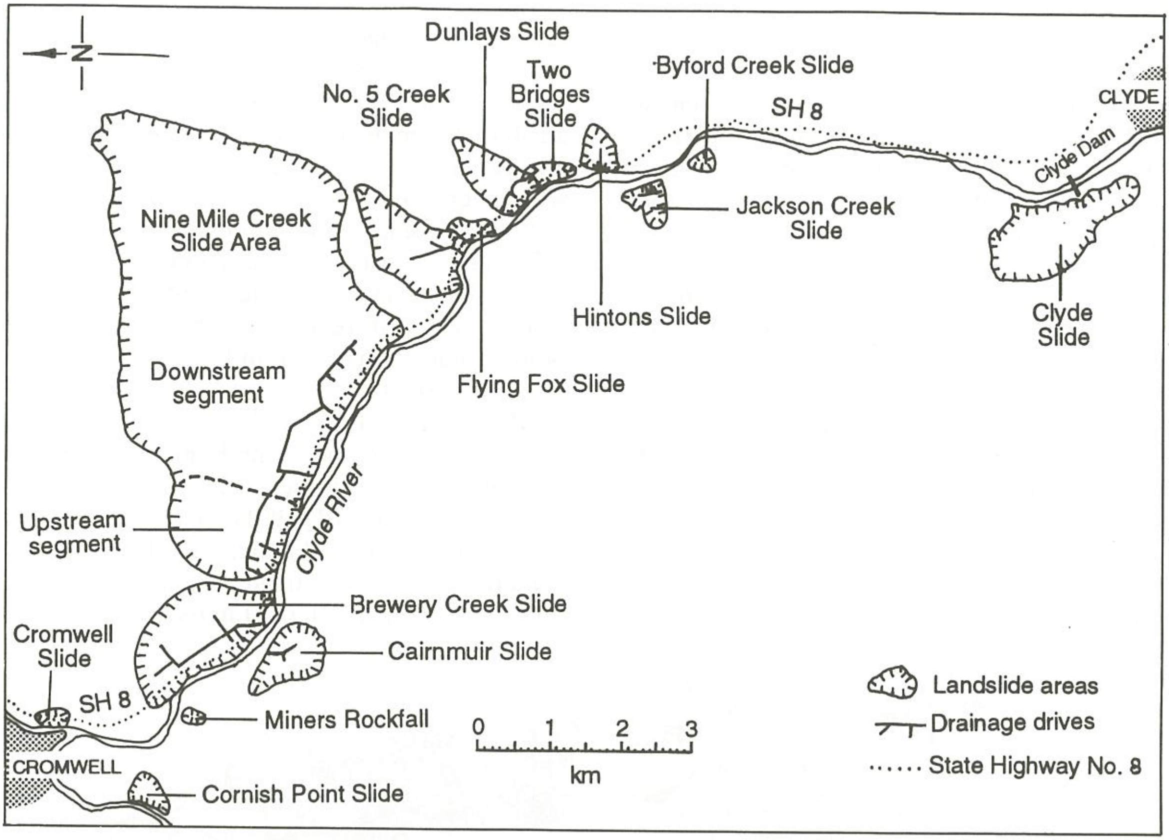

Figure 20.16 Map of landslides in Cromwell Gorge, Central Otago. Over 25% of the shoreline of Lake Dunstan, behind the Clyde Dam, consists of landslide material in the gorge. The toes of some slides will be flooded to a depth of 60 m (After Gillon and Hancox, 1991).

results from lower resistance at the toe, because submerging the toe reduces its weight and therefore lateral support. Increased groundwater pressures reduce the effective stresses and therefore strength along the shear plane. The exact impact varies according to the geology of the slide, the amount of lake rise, and the nature of the groundwater system. In Cromwell Gorge the slides, or parts of them, will undergo reductions in stability following the filling of Lake Dunstan ranging from 2-20% (Gillon and Hancox, 1991). Because of the critical control groundwater has on the stability of these slides, drainage measures are the most effective and economical means of stabilisation.

Stabilisation work has therefore focused on two main methods of increasing strength and reducing the impact of lake filling on the stability of the slopes - buttressing and drainage. Buttressing involves compacting rock and gravel at the base of the slide areas to add weight and therefore lateral support and strength to the toes. Groundwater conditions have been controlled through a series of tunnels excavated into the slide areas to provide permanent drainage. Because of the compartmentalised nature of the groundwater within the slides, a fan of drill holes have then been drilled from the tunnels to lower the groundwater over the whole slide area. This stabilisation work has required 14.5 km of tunnels, 60 km of surface drilling and 78 km of drainage holes drilled from within the tunnels. Five million cubic metres of buttressing material has been placed at the base of slide areas.

Cairnmuir Slide

Cairnmuir Slide provides a good illustration of the problems in the Cromwell Gorge and the impact of stabilisation techniques (Figures 21.16 & 21.17). This slide, on a 20° slope, is currently moving at a rate of 10-100 mm per year and shows a total displacement of 600 m since the initial failure occurred (Gillon *et al.*, 1991).

The groundwater in the slide area belongs to two systems; a perched water table within the mobile material and a sub-basal system below the shear plane. The perched water table and resulting pore-water pressures above the shear plane have been shown to control the movement of the slide through their influence on the effective stresses acting on the shear plane. Since natural drainage of this saturated material is through the shear plane, the sub-basal system also indirectly controls stability.

During July and August 1990 the slide accelerated (to a maximum of 0.8 mm/day) resulting in a total of 16 mm downslope displacement over about 6 weeks (Figure 21.18). This movement was attributed to high infiltration from increased rainfall during the preceding four months, coupled with lower evaporation because of the time of the year, thus raising groundwater levels. A "theoretical" water balance analysis (supported by field tests) which considered the inputs, outputs and storage of moisture in the system was used to predict changes in the groundwater in response to climatic variables. The modelled groundwater conditions correlated well with movement, although at the time no groundwater response was recorded in the few piezometers that had been installed (Gillon *et al.*, 1991). The water balance model provides a good indication of groundwater conditions, but overestimates the actual response e.g., the model predicted a 1 m rise in the perched

Figure 20.17 Aerial photograph of the Cairnmuir landslide in Cromwell Gorge. (Photograph: Courtesy of the Clyde Power Project, Electricity Corporation of New Zealand)

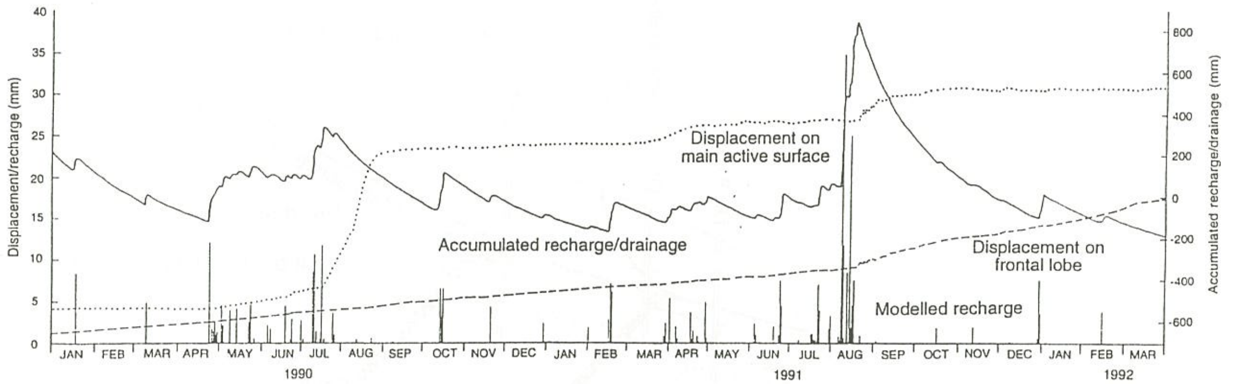

Figure 20.18 The water balance model developed for the Cairnmuir landslide in Cromwell Gorge shows how groundwater conditions affect movement rates. Daily precipitation, evaporation, and infiltration were used to model groundwater recharge. Prior to draining the slide, peaks in modelled groundwater conditions correlated with accelerated movement on the main active surface. Following drainage both total movement rates, and the influence of groundwater conditions, appear to have been significantly reduced. (Data supplied by the Clyde Power Project, Electricity Corporation of New Zealand)

water table when only 350 mm was actually recorded. This overestimation may result from applying a general model to a specific site with a highly compartmentalised groundwater system, or because, as has subsequently been shown, the groundwater reacts more to infiltration through tension cracks than to general infiltration (Lilley, 1992 *pers com.*).

Investigations have therefore been directed towards identifying groundwater systems within and beneath the slide and the feasibility of draining these to improve stability. The drainage measures adopted are similar to those being undertaken on several other landslides in the area. A Y-shaped tunnel was driven beneath the slide from which a "fan" of drainage holes were drilled. One link of this tunnel was driven behind the frontal lobe while the other was pushed straight up the slide. These tunnels were all at least 30 m below the active failure surface. Drainage drilling in the perched water table also has been undertaken. The project involved 600 m of drive (3.5 m in diameter) and 6.5 km of drilling from this drive. Approximately two-thirds of the drilling was into the perched water table and one-third into the sub-basal groundwater. These measures will drain the confined sub-basal groundwater and control the rise of groundwater when the lake is filled (Figure 21.19). The improvement achieved in stability over existing conditions is likely to be small due to difficulties in draining the perched water table. An attempt has also been made to reduce infiltration into the slide aquifers by regrading stream channels on the slide, and lining these channels where they cross sink hole systems and areas of high erodibility (Gillon *et al.*, 1991)

The reduction in groundwater levels (drawdown) resulting from this work has been close to that predicted by the computer models (between 10-40 m) especially in the sub-basal zone where the groundwater system is simpler. The response of the sub-basal groundwater to the two phases of the drainage programme are illustrated by the drawdown curves (Figure 21.20). The initial drawdown is a response to the large drainage drive and the subsequent drop shows the effect of the fan drilling from within the drive which was designed to drain a wider area. While draining the sub-basal water table has no direct effect on stability, by increasing the hydraulic gradient through the shear plane the drainage of the perched water table will be increased and will improve the factor of safety by about 0.5%. The perched water table is now less than 5 m above the shear surface when it was

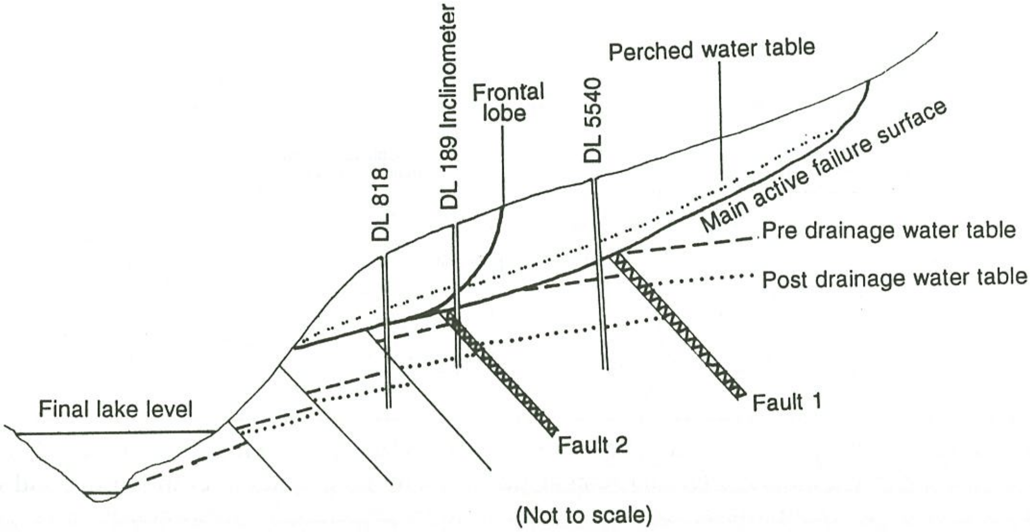

Figure 20.19 Diagrammatic cross-section of the Cairnmuir landslide, showing the main structural components and groundwater compartmentalisation. Tunnels have been drilled to lower both the perched and sub-basal water tables. (From a sketch supplied by the Clyde Power Project, Electricity Corporation of New Zealand)

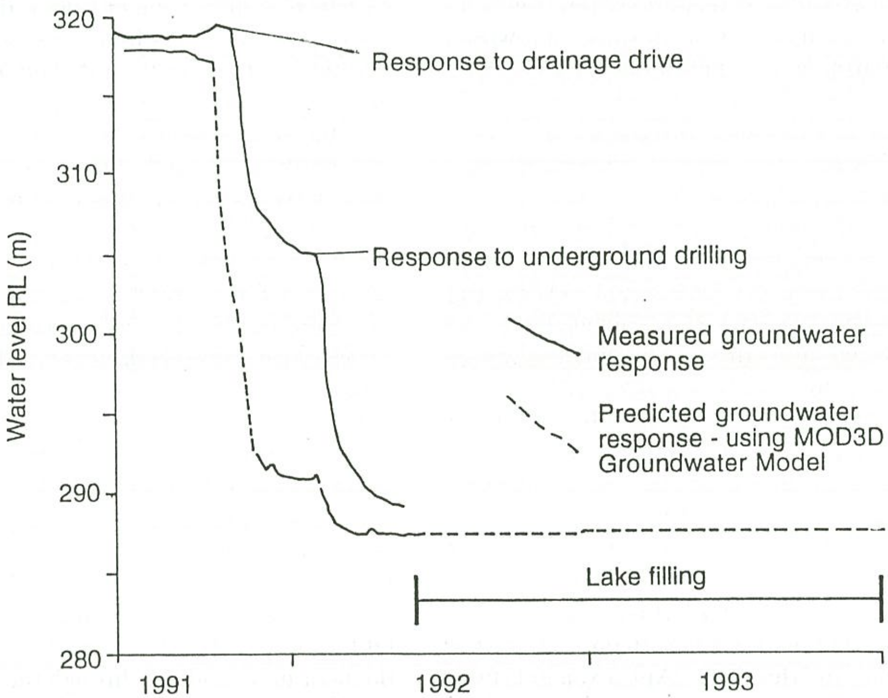

Figure 20.20 Changes in the sub-basal groundwater level within the Cairnmuir landslide. The initial drop in levels (piezometer DL540) followed construction of the drainage drive, and the subsequent drop reflects the impact of fan drilling from within the drainage drive. (From data supplied by the Clyde Power Project, Electricity Corporation of New Zealand)

initially 18 m, producing a 3.5% increase in the factor of safety. While this improvement in stability is small, a drop in the factor of safety of only 0.3-0.5% resulted in an acceleration of movement of the slide of more than eight times. To show how delicate the system is with respect to stability when a small berm (5 m x 1.5 m) was cut at the base of the slope movement accelerated five times in this area.

The work undertaken to stabilise the landslides in the Cromwell Gorge has shown the importance of the groundwater in controlling stability, how short-term changes in groundwater levels can affect the movement rates, and how the most cost-effective means of improving the stability of the slopes and minimising the impact of filling Lake Dunstan is through draining the pore-water.

Summary

Water content is a critical variable in slope stability; it changes rapidly and this leads to variation in strength over a short period of time. Through its effect on the cohesive and frictional strength of the material, water is a common trigger of slope instability, as can be seen in the high correlation between wet conditions and slope failures.

By understanding the role of water in slope stability we can manage our use of the landscape to minimise the risk, while maximising our development options. No matter how careful our management strategies, however, nature will continue to impose a "tax" from time to time for our use of New Zealand's dynamic landscape.

Acknowledgements

I am most grateful to Dr Mike Crozier for providing constructive criticism of this material and John McKinnon and David Winchester for ensuring that it was intelligible. I would also like to thank the Electricity Corporation of New Zealand Ltd, Works Consultancy, and in particular Peter Lilley of Riley Consultants Ltd for access to and permission to publish recent data on the Cairnmuir landslide investigation and stabilisation project. Financial assistance for the research in the Wairarapa discussed in this paper was provided by the National Water and Soil Conservation Authority. Heather Campbell, supported with funds provided by the Internal Grants Committee of Victoria University, drafted many of the diagrams.

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21
Hydrology and Large Water Projects

H J Freestone

Introduction

Many large and intricate engineering works have been constructed in and around New Zealand rivers. The most significant are dams and the lakes (reservoirs) they create.

Hydrological analysis associated with the design and operation of dams in New Zealand is extensive and so the focus of this chapter is dams and in particular hydro-power dams and associated works. Hydro-power development in New Zealand provided the main impetus for early hydrological data collection programmes in New Zealand (chapters 1, 3).

Dams in New Zealand

Dams in New Zealand have been built for storage of irrigation water, hydro-electric power generation, domestic and industrial water supply, and floodwater storage. The New Zealand dam inventory (Ministry of Commerce, 1988) lists 402 dams which are over 1.8 m high and which have water storage capacities greater than 18,500 m^3. A few dams will have been missed from the survey, as the list does not include thousands of small farm water-supply dams and historic dams of the past such as the notable kauri dams used last century. Details of size and capacity for forty large New Zealand dams are listed in Table 21.1 (NZ Society on Large Dams, 1989).

The highest dam in New Zealand is Benmore, which is 118 metres from river bed to dam crest. Maraetai and Matahina dams, the highest in the North Island, are 87 and 86 metres respectively. The greatest reservoir capacity behind a dam in New Zealand is the 6,300 million cubic metres of water stored behind the Pukaki dam.

Dams and the Flow Regime

Dams have significantly altered the hydrological regime of many New Zealand rivers. Water supply dams impound all the water from the catchments upstream, apart from excess flood water. The impounded water is piped away, and much of it does not return to the river of origin, leaving depleted flows in channels downstream.

Some large reservoirs behind dams modify flow regimes downstream by reducing flood peaks or changing patterns of seasonal flow (Figure 21.1). In Lake Pukaki, periods of high inflow from November to February are used to collect water. This stored water is then used in winter when there is a high demand for power in New Zealand. In the Pukaki catchment winter is generally a period of low inflows to the lake, so that high energy demand is met by the seasonal transfer of water using a large reservoir system. In Lake Pukaki the storage volume which is available for use is 1990 million cubic metres (Freestone and Maslin, unpublished report, 1992).

Dam	River	Dam Height (metres)	Reservoir Capacity (m^3 x 1000)	Spillway Capacity (m^3/s)
Waingaroa	Waiwhakangarongoro	33	5100	6
Whau Valley	Waiarohia	26	2045	195
Lower Nihotupu	Nihotupu	24.7	4805	480
Waitakere	Waitakere	25.3	1850	-
Hays Stream	Hays Stream	25.9	1230	100
Wairoa	Wairoa	47	12000	122
Mangatangi	Mangatangi	78	37000	510
Kapukapu	Kapukapu Stream	15	140	47
McLaren Falls	Mangapapa	23	250	1600
Mangaonui	Mangaonui	29	370	125
Matahina	Rangitaiki	86	33000	1980
Karapiro	Waikato	66.8	87000	849
Arapuni	Waikato	64	33300	778
Waipapa	Waikato	37	11400	880
Aniwhenua	Rangitaiki	10	1600	1270
Maraetai	Waikato	87	94200	850
Whakamaru	Waikato	56	80000	730
Aratiatia	Waikato	17.8	-	800
Hinemaiaia	Hinemaiaia	21	-	85
Rangipo	Tongariro	23	456	1560
Mangamahoe	Mangamahoe	26	1000	150
Waikaremoana	Waikare	*	58000	91
Moawhango	Moawhango	68	113000	1100
Patea	Patea	83	144000	3880
Makara No1	Makara Stream	12	835	132
Tiritea	Tiritea	39	1682	140
Mangahao	Mangahao	31.7	1560	1270
Cobb	Cobb	32.7	28000	950
Maitai	Maitai	39	4000	280
Waihopai	Waihopai	35	0	700
Lake Argyle	Branch	15	1875	2500
Pukaki	Pukaki	61	6300000	3968
Ruataniwha	Ohau	43	50000	1740
Benmore	Waitaki	118	2100000	3400
Hawea	Hawea	32	2180000	360
Aviemore	Waitaki	56	425000	4250
Waitaki	Waitaki	47.8	53200	5800
Clyde	Clutha	102	350000	5500
Roxburgh	Clutha	76	100000	4250
Mahinerangi	Waipori	39	220	180

* natural barrier

Table 21.1 Details of 40 New Zealand dams (Source NZ Society on Large Dams "Dams in NZ", August 1989)

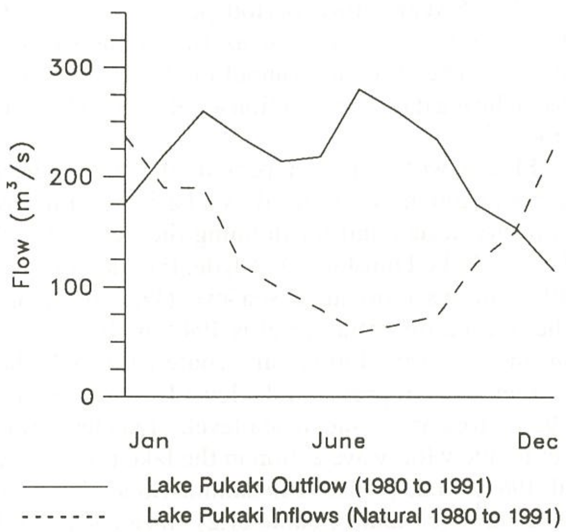

Figure 21.1 Lake Pukaki inflows and outflows 1980 to 1991

Large Scale Hydro-electric Development

New Zealand generates 75% of its power from hydro-electric power stations on its rivers and lakes (Ministry of Commerce, 1991). Water is abundant and there are a large number of fast flowing comparatively steep rivers. Dam sites need a combination of sufficient flow and fall to produce energy. Other requisites are sound bedrock for foundations, narrow gorges, and minimum disruption to the upstream environment from the lake formed by the dam.

The Clutha River

In the past, particularly in the late 1950s through the 1970s, dams and power stations were completed in quick succession to meet increased power demand. Since then development has slowed, allowing hydrologists more time to prepare comprehensive hydrology studies. Such a study has been made for New Zealand's largest catchment, the Clutha, where there are many different options for hydro-power development (Jowett and Thompson, 1977).

Part I of the study covered rainfall, flow, rainfall-flow relationships and flood estimates. Part II was a detailed analysis of basic water level, flow and rainfall records (Jowett and Thompson, 1977). Although the study comprehensively covered major tributaries to the Clutha River, it focused attention on dam sites being considered for development at the time (Table 21.2). The site labelled DG3 (Clyde) refers to the Clyde Dam which began generation of electricity in 1992.

The Roxburgh dam and power station were completed in 1956. In 1959 a dam was completed at the outlet of Lake Hawea. This structure had no power station; it was built to provide storage for

Station	Mean Flow	Peak discharge (m^3/s)			
		15 year	500 year	1000 year	PM Flood
Hawea Lake Control	63	0	316	316	316
Wanaka Lake Outlet	190	750	1210	1300	2205
Wakatipu Lake Outlet	156	570	909	1050	1661
Upper Pisa Dam Site	260	750		1610	2587
UC8 Dam Site	261	750		1610	2587
K7/2 Dam Site	200	950	1730		2400
K8 Dam Site	211	1020	1930		2760
DG3 (Clyde) Dam Site	**472**	**1800**	**3200**		**6820**
Roxburgh Power Station	490	2000	3600		7830

Table 21.2 Design discharges for existing and proposed dam sites on the Upper Clutha River. PM Flood is the probable maximum flood. From Ministry of Works report - Clutha Power Development - "Flows and Design Floods, Jowett and Thompson, 1977.

the power station 110 km downstream. The dam raised Lake Hawea's level by about 19 metres and created an extra 2,400 million cubic metres of storage. Some of this extra storage is no longer used for power generation. Whenever the lake was drawn down to low levels, wind picked up the exposed dried lake sediment along the shores, creating dust storms, so restrictions were imposed on lake lowering. Storage available for power generation is now 1,100 million cubic metres but the extra storage can be used under extreme circumstances.

Stored water is used to meet electricity peak demands, for example, the increased demand for power at around 5:30 pm when New Zealanders turn on electric stoves, television sets and other electric appliances as part of their evening meal and relaxation time. Water is also stored for use in other seasons. This is illustrated by Lake Pukaki data (Figure 21.1).

Clyde Dam

The Clyde dam is 102 metres high; its reservoir is known as Lake Dunstan. At normal operating lake level it has a generation head of 60 metres. Generating or nett head is defined as the difference between head water level and tailwater level, but it can be more simply considered to be the difference in height between the water level at the top of the dam and the water level in the river immediately downstream of the powerhouse.

Reports such as the 1977 Clutha hydrology report provide more than a series of hydrological facts and figures. A study of various peak discharges at the Clyde Dam site (Table 21.2) illustrates the use of flow values for design. The mean flow of 472 m^3/s is a general indicator of capacity for power generation, but if adjusted by subtracting estimates of spillflow it can be used with the nett head to accurately calculate annual power station generation. At Roxburgh Dam, downstream of Clyde, water available for power generation (1966-1991) was 83% of mean flow. A further 17% was flood spill water. For the Clyde dam predicted annual generation is 1,930 (GWH). One GWH is equal to a thousand million watts of power supplied over 1 hour.

The 15 year return period peak flow of 1,800 m^3/s for Clyde is known as the "construction flood". The diversion tunnel or diversion works used during dam construction are designed for this flood size.

Floods with a return period of five hundred years (3,200 m^3/s) are used as a basis for spillway capacity design and for defining the design flood level in Lake Dunstan. At Clyde, the dam crest is 197.1 metres above mean sea level (Dunedin), and the normal operating level is 194.5 metres above mean sea level. Floods are routed through the system so as to prevent lake level from exceeding 195 metres above mean sea level. This level was set to allow for wave action in the lake (Hatton et al, 1987). The probable maximum flood (PMF) is 6,820 m^3/s (column 6, Table 21.2) - it can be passed by using all available sluice and spillway capacity at the dam.

Clyde Dam is a concrete dam and so a 500 year return period flood was used for flood design. Other sites on the Clutha River where an earth or rock fill dam may have been contemplated had estimates of the 1000 year return period flood (Table 21.2).

Setting levels for the dam crest and the maximum operating water level, working out the number of spillway gates needed and their dimensions, calculating the number of water turbines and electrical generators required is all part of the design process. This process also considers construction costs for various dam heights and limits dictated by foundation conditions. The final design will often involve a range of options.

As well as being used in design calculations the flood size data are often used by physical modellers who build scale models of dams. Scale models are used to test various design options in a hydraulics laboratory. In New Zealand modelling facilities are located at the Central Laboratories (Works Consultancy Services, Gracefield) and at Auckland, Canterbury and Lincoln universities.

A scale model of the Clyde dam was used to test and modify sluice gate, spillway and tailwater channel design.

Detailed Analysis

Summary data as presented in Table 21.2 are only a small fraction of the data presented in a detailed hydrological study such as that carried out for the upper Clutha River.

Seasonal water availability is modelled in conjunction with lake storage capability to produce seasonal lake outflows available for power generation.

Daily generation patterns at Roxburgh Power Station were available for Clyde design. The relationship to Roxburgh generation and the use of the Lake Hawea storage facility is illustrated in Figure 21.2. Daily power generation cycles are illustrated by flows at Roxburgh Dam. The magnitude and amplitude of the flow peaks and cycles increase when flows from Lake Hawea are increased, illustrating how Lake Hawea can be used to boost power generation.

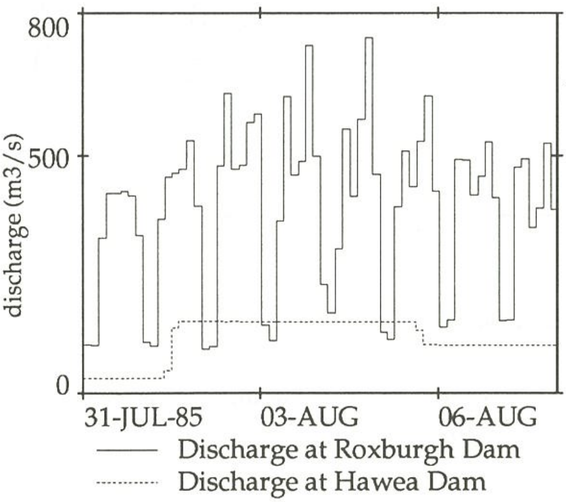

Figure 21.2 An example of Lake Hawea flow releases and generation at Roxburgh Power Station

Historic Floods

Information on historic floods can provide useful additions to available data on river flow and rainfall. For example, historic flood levels for the 1878 floods at Alexandra were higher than those calculated by backwater analysis for the design flood for Lake Roxburgh. The higher levels were used for fixing the design flood level at Alexandra, not just because they were higher, but for a variety of reasons, such as future sedimentation effects.

Sedimentation

The design life of a reservoir is calculated by estimating the rate at which sediment enters the reservoir, and adjusting for the lake's trapping efficiency - the percentage of incoming sediment retained by the lake. The effects of reservoir sedimentation after 30 and 100 years of operation of the Clyde Dam have been calculated, and used to produce inundation maps which were a basis for the purchase of land around the perimeter of Lake Dunstan. Sediment studies were carried out subsequent to the 1977 report (Jowett and Thompson, 1977), including special collection of sediment data to identify source rivers. Lake Dunstan bifurcates 19 km upstream of the dam; one branch is known as the Clutha arm and the other the Kawarau arm. The greatest sediment input comes from the Shotover River which brings its sediment load via the Kawarau River. Sediment input to the Clutha arm is only 8% of the total sediment entering the reservoir, compared with 92% entering via the Kawarau arm.

Data on reservoir sedimentation have been analysed for Lake Roxburgh downstream of the Clyde Dam (Figure 21.3). The rate of accumulation of sediment in the lake is obtained by comparing lake depths in 1971 with those in 1989 (the lines represent lake bed level as defined by maximum depths).

Average sediment accumulation in Lake Roxburgh as calculated in 1990 is 1.5 million cubic metres per year. This is based on regular surveys at approximately five year intervals. Over a 28 year period the volume of Lake Roxburgh has been reduced by 42%. This has had little effect on the normal operating range of Lake Roxburgh.

By comparing calculated sediment input to Lake Roxburgh with sediment deposition in the lake it is possible to estimate "trap efficiency" for

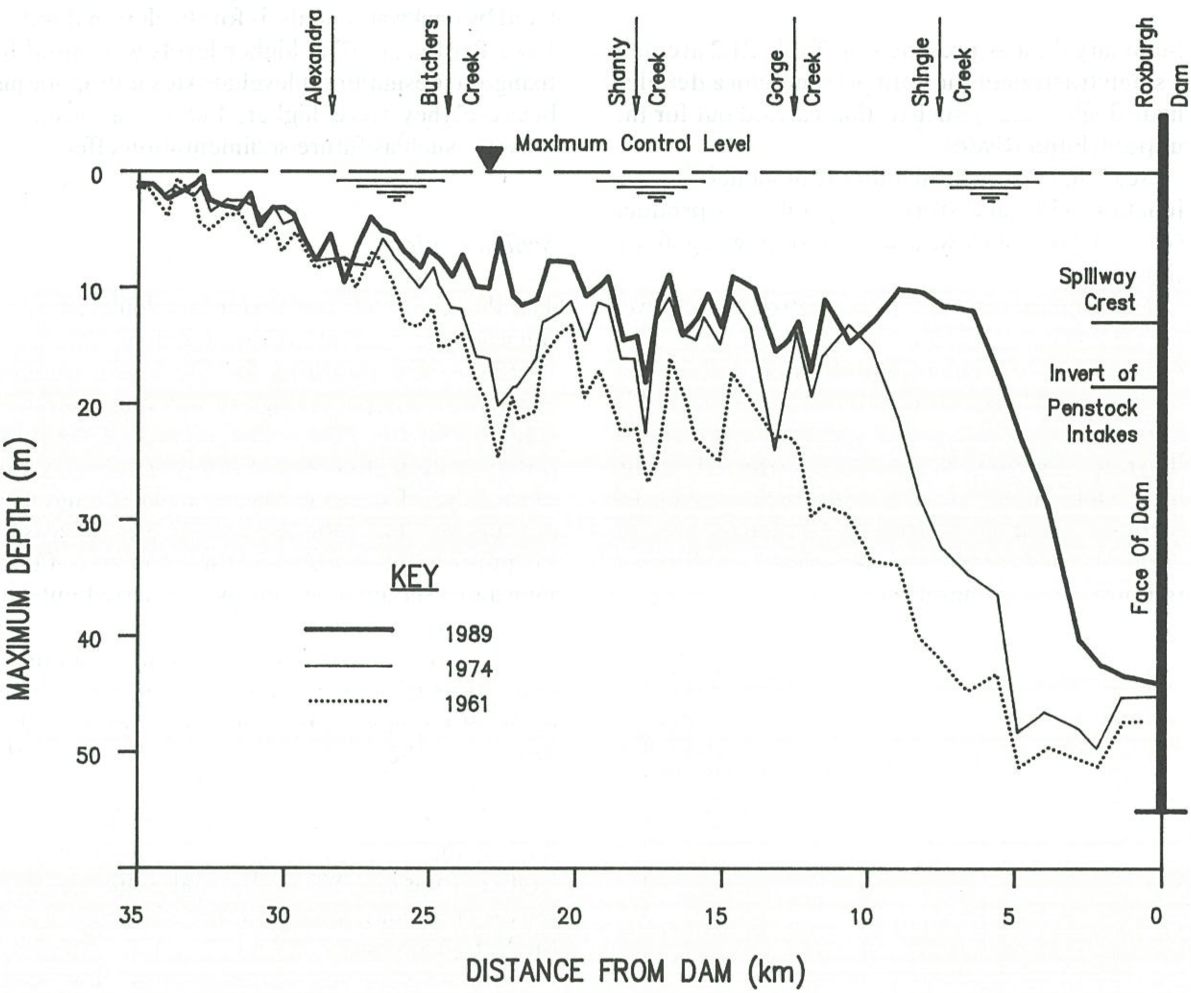

Figure 21.3 Sediment levels in Lake Roxburgh 1961 to 1989, based on soundings. (From Works Consultancy Services report "Lake Roxburgh Sedimentation, 1990).

the lake. Trap efficiency for Lake Roxburgh is estimated to be 80%, that is 80% of the sediment entering the lake remains in it (Jowett and Hicks, 1981). Sediment information from Lake Roxburgh has been used for estimating sedimentation rates for Lake Dunstan upstream of the Clyde dam, and for estimating lake-bed sedimentation profiles for backwater modelling in the lake. Backwater studies have in turn been used to predict flood levels after 30 and 100 years of sedimentation in Lake Dunstan.

Lake Filling

Several other studies have been carried out in connection with the Clyde dam. For example, work on the landslides upstream of the dam indicated that particular care would be necessary during the lake-filling stage of dam commissioning. The level of the lake was to be raised to various holding levels where the water level was to be maintained with minimum variation. Extensive modelling was used to define operating practice, particularly during times of floods.

Seasonal flood patterns are quite marked in the Clutha catchment and so consideration was also given to timing for lake-filling. Flood risk is lowest in June and greatest in January and March (Figure 21.4).

Operational Hydrology

Lakes Manapouri and Te Anau - Guidelines

A significant aspect of hydrological activity associated with dams relates to operating practices. For Manapouri power scheme, most operational practice is defined by guidelines gazetted by the Government in 1981, after consultations between the Crown and the Guardians of Manapouri and Te Anau (New Zealand Government 1981). Revisions in 1990 incorporate changes agreed to by the Guardians and the Electricity Corporation of New Zealand (New Zealand Government, 1990).

The essence of the guidelines is to maintain a natural lake level regime where possible, particularly for high and low lake conditions. They recognise three operating ranges for Lakes Manapouri and Te Anau - the high, main and low operating ranges. For Manapouri the main operating range is 1.8 metres and for Lake Te Anau 1.3 metres (Table 21.3).

Thirty five years of records (1934-69) were used to establish natural main, low and high operating ranges. For the 1990 revision simulated records and a period of pre-1934 record for Lake Te Anau were also considered (Works Consultancy Services, 1989).

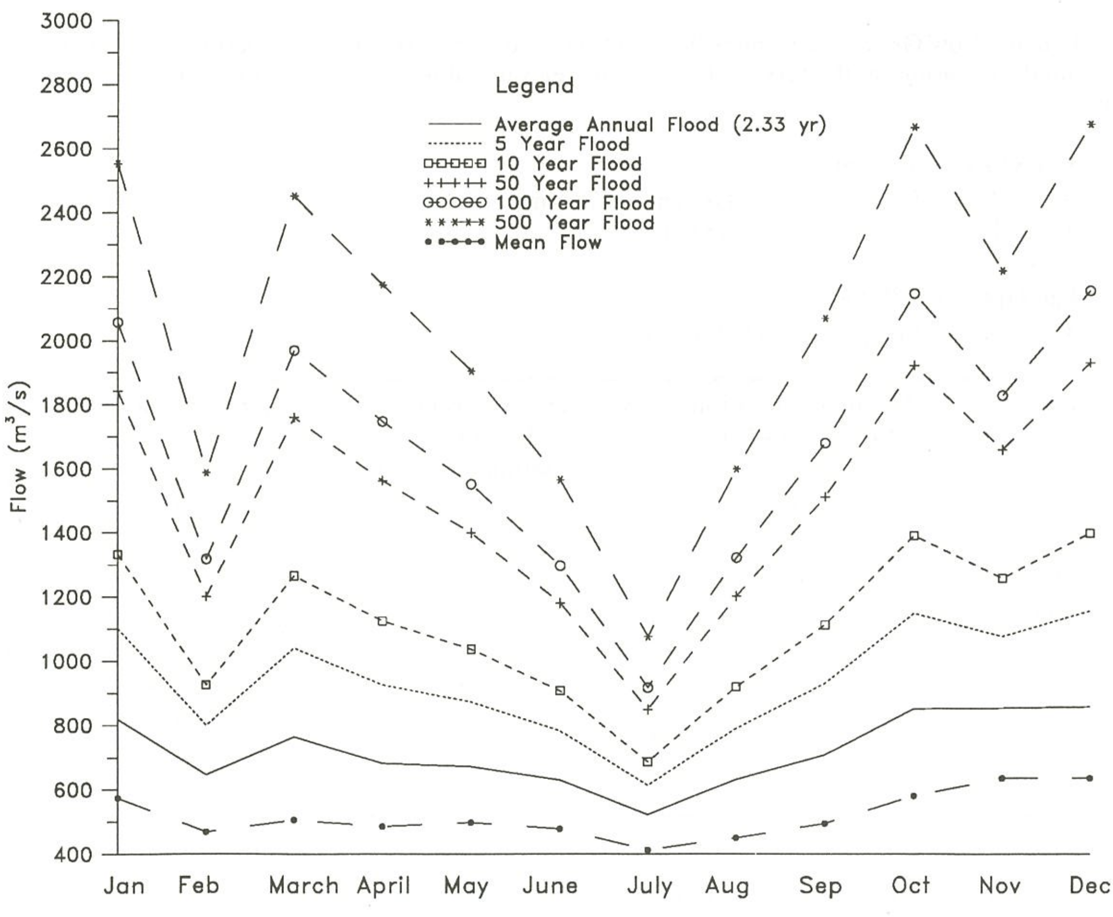

Figure 21.4 Clutha River at Clyde - Seasonal variations for a range of flood sizes

Manapouri-Te Anau Development Act 1963

Manapouri-Te Anau Development Act 1963
Pursuant to section 4A of the Manapouri-Te Anau Development Act 1963, the Minister of Energy, based upon the recommendation of the Guardians of Lake Manapouri and Te Anau and the Electricity Corporation of New Zealand Limited, hereby gives notice that the following guidelines shall apply to the operations, for hydro-electric purposes of Lake Manapouri and Te Anau and the associated power station.

These guidelines may be subject to review after a period of 1 year.

1. Lake Management
The parties recognise three operating ranges for each lake, namely:

Main, High and Low as set out in the tables below.

In each range the Electricity Corporation may operate within the levels set out, but will endeavour to maintain continuous variations within that range.

In the High and Low Operating Ranges the Electricity Corporation will use its best endeavours to stay within the durations and intervals shown in the relevant table at the levels set out in the tables below.

(a) Main Operating Range

(i) Lake Manapouri from 176.8 m to 178.6 m.
(ii) Lake Te Anau from 201.5 m to 202.7 m.

(b) High Operating Range

(i) Lake Manapouri above 178.6 m.

Elevation (m)	Maximum duration (Continuous Days	Minimum interval between floods in this level (Continuous days)	Interval/duration Ratio
At 180.5	1	100	100.0
Above 180.4	3	100	33.0
Above 180.1	9	100	11.0
Above 179.8	22	80	3.6
Above 179.5	35	40	1.1
Above 179.2	44	40	0.9
Above 178.9	99	20	0.2
Above 178.6	119	20	0.2

Table 21.3 Extract from Manapouri - Te Anau Gazette Guidelines (1990)

Low Operating Range

Guidelines for minimum lake levels consist of a series of steps defined by a threshold lake level and a duration in days. For example, Lake Manapouri may stay below lake level 176.2 metres above sea level for only 35 days during one single low level-event. Furthermore, there are to be no more than two "events" in a year for each step. Durations were derived mainly from the three worst droughts.

High Operating Range

When the guidelines were revisited in 1990 no revision was made of the guidelines for maximum allowable lake levels, in spite of their severe testing in 1988. High-range guidelines have worked well for most flood conditions. The 1988 flood was the highest in 120 years for both lakes and it is recognised that guidelines do not generally cover such extreme events (Freestone 1988). However the 1.9 metre high range for Lake Manapouri and the 1.6 metre range for Lake Te Anau cover most situations. Range steps are defined in a similar way to the low range steps. For example, Lake Manapouri should not remain above a level of 180.1 metres above mean sea level Deep Cove for more than 9 days at time. Furthermore, a minimum interval of 100 days is required between floods. Within the guideline table (Table 21.3) there are further restrictions defined by an "interval/duration guideline ratio".

General Guidelines

Several other guidelines have been derived to assist with "natural" operation of both Lakes Manapouri and Te Anau.

Best endeavours must be used to avoid drawing Lake Manapouri below 176.2 metres during the equinoxial periods (March, April, October and November).

When drawing down the level of the lakes the drawdown rate should not exceed 0.05 metres per day for Lake Manapouri and 0.03 metres per day for Lake Te Anau, averaged over four days. Guidelines are also designed to keep the mean annual lake level within the main operating range.

Level guidelines based on the five-year running mean have been difficult to maintain and were deleted from the 1990 guidelines.

Monitoring

Gazette guidelines are monitored each year with a report to the appropriate government Minister. This mandatory report requires that certain basic records be kept. The most critical of these are lake levels and total inflows and outflows for both Lakes Manapouri and Te Anau.

Other checks are made from time to time, including sedimentation in the Lake Manapouri outlet channel, beach profiles and checks on Lake Te Anau control gate operations relative to fisheries.

At present Lake Manapouri beach profiles are surveyed periodically. Recently a comparative programme was set up on Lake Hauroko whereby beach profile changes for both lakes could be compared to hydrological parameters such as lake range, the time each lake spends in its low operating range and other relevant characteristics.

Operating practises can change from time to time to correct obvious problems. For example, it was found that holding the lakes at a constant level for too long created shoreline "benching". Variable levels have been introduced to avoid such problems. In general the guidelines have a "best endeavour" clause which, although sometimes hard to define, is necessary to cover the extremes of flood and drought.

Ongoing Hydrological Analysis

Since 1967 there has been a statutory system for monitoring and managing the use of water in New Zealand. Recently this system has been revised, and is now covered under the comprehensive Resource Management Act.

Large hydro-electric systems such as the Waitaki, Waikato, Manapouri and Clutha power developments usually involve more than one river, dam or major control structure. These systems have complex management requirements which are usually based on detailed hydrological analysis.

Waitaki River

Water rights were recently renewed for the eight Waitaki River power stations and associated structures (Freestone 1990, Canterbury Regional Council, 1990). One of several new conditions is a minimum flow for the Waitaki power station, which is the station furthest downstream. A flow of 120 cubic metres per second was set as the minimum after looking at user requirements and past hydrological records. The flow of 120 cubic metres per second could be reduced under extreme drought conditions. Provision needs to be made to establish when a natural drought is occurring in a managed catchment such as the Waitaki. This requires careful simulation modelling. Even then there is a degree of scepticism because the answer is calculated rather than measured.

The Waitaki water rights also stipulate an eight to twelve cubic metres per second flow release from Lake Ohau for fisheries and irrigation. Although a large part of the water bypasses only one power station (Ohau 'A') it does represent a loss in power generation. A concession was gained in the operation of Lake Ohau to offset to a small degree the water "given away" in the public water allocation. However, for the hydrologist and power station operators, a challenge is issued. Can different strategies for operating other parts of the 8 power station system and associated storages, such as lakes Tekapo and Pukaki, be used to offset generation loses caused by the Lake Ohau flow releases?

The answer is a cautious yes, but it involves a canal system modification which means more capital expenditure. It also requires hydrological modelling of not only the Waitaki River power stations but all parts of the national power generation system. Provisional modelling based on past records (1981-90) indicate that flood water previously spilt can be used to generate morepower at Ohau 'A' power station provided a critical canal structure is modified.

Waikato River

Recent upgrading of the spillway of the Arapuni dam has allowed the eight power stations of the Waikato River system to revert to an operating regime last used in 1975. The Waikato Regional Council had to consider the matter and allow for the public to become involved. Two technical reports were produced and the Waikato flood rules were upgraded (Works Consultancy Services, August 1990). The rules were based on design flood routing through the whole system. When the work was presented to six separate public meetings it became obvious that the public were not really interested in the design flood but in the flood with a return period of once in five years. It is this flood that regularly influences the lives of people, particularly in the lower Waikato region. Great care must be taken when minimising the downstream effects of smaller floods to ensure that flood storage capacity held in the dam reservoir for the large design flood is not used up or compromised. It is this design flood capacity which is essential to the safety of the dam structures.

Mangahao 'Number 2 ' Dam

The Mangahao power station is near Palmerston North and the scheme consists of three dams. One of these, 'Number 2' dam needs annual desilting so that the power station can operate efficiently.

Studies on the effects of desilting have been carried out in 1982 and 1992, addressing both water quality and aquatic issues (Manawatu Catchment Board and Regional Council, 1982). The 1992 study also involved a trial by canoeists to see if canoeing was possible during desilting flow releases (Works Consultancy Services, 1992). The reports form a database from which future management studies can be made.

An extract from the 1992 Mangahao desiltation study gives water and sediment volumes for eight separate flow releases over two days (Table 21.4). The two measuring sites on the Mangahao River at 'Number 2' Dam and Kakariki are 15 km apart. Measurements were also made of dissolved oxygen, water temperature, optical water clarity, organic matter, nutrients and fish population.

Flush Period		Peak flow (m^3/s)	Total released Flow (m^3)	Mean Sediment Concentration (mg/l)	Total Sediment Load (tonnes)
Start	End				
					25 March 1992
0945	1020	30.2	46030	434	20
1205	1510	37.6	119000	479	57
1640	2000	40.5	135000	429	58
					26 March 1992
1002	1124	61.1	289000	500	145
1232	1315	33.9	88000	454	40
1423	1450	45.1	95000	495	47
1521	1557	36.7	77000	480	37
1615	1650	36.1	76000	474	36

(a) Number 2 Dam

Flush Period (at No. 2 Dam)		Peak flow (m^3/s)	Total (m^3)	Mean Sediment Concentration (mg/l)	Total Sediment Load (tonnes)
Start	End				
					25 March 1992
0945	1020	8.3	32200	124	4
1205	1510	18.5	96500	238	23
1640	2000	16.2	108400	175	19
					26 March 1992
1002	1124	38.7	173300	548	95
1232	1315	19.8	64800	308	20
1423	1450	23.8	39000	435	17
1521	1557	26.4	71900	445	32
1615	1650	28.0	63700	487	31

(b) Kakariki

Table 21.4 Mangahao River - Water and sediment discharges during flow releases to desilt "Number 2" dam, measured at (a) Number 2 dam and (b) Kakariki, 15 km apart.

Conclusion

Today large-scale engineering works associated with rivers must satisfy a variety of Government regulations. Although much of this is not new, there are more stringent environmental requirements now than there were in the 1950s and 1960s when many large developments were either built or planned.

Dam safety and water management requirements make ongoing hydrological design and planning essential. Simulation and hydrological assessment requires good data collection and archiving systems. Although the amount of hydro-electric development may have slowed in New Zealand the requirement for ongoing hydrological assessment has not diminished.

At the present time, (November 1992), a Prime Ministerial Review of the 1992 winter power shortage is in process. Because of this, it is not appropriate to present a detailed analysis of the power shortage here. It is reasonable, however, to present some brief details of hydrological conditions leading up to the power shortage. Material quoted here was presented at a public meting held in Timaru on 10 July 1992.

Power shortages became critical when key South Island hydro-storage lake levels were very low. Two of the lakes, Tekapo and Pukaki provide over 60% of the national hydro-storage.

For the seven months, November 1991 to May 1992, inflows to Lakes Tekapo, Pukaki, Ohau, Hawea and Wanaka were the lowest collectively in 62 years of recording (Figure 21.5, Freestone 1992). All five lakes individually recorded the lowest inflows during the period except for Lake Tekapo, where the second lowest inflows in 67 years were recorded. There was only a 1.45% chance that inflows for the five lakes would be as low as they were for the seven months ending May 1992.

The seven month period is particularly critical because 73% of the annual inflows are usually received during this time.

However, not all assessments show the event to be as extreme as the inflow study suggests. For example, a rainfall analysis for the same period at the Hermitage, located at the head of Lake Pukaki, shows a return period of once in 10 to 15 years. This compares to the inflows where the return period is approximately once in 60 to 70 years.

Many reasons for these differences are possible. The processes whereby precipitation moves through the hydrological system, to appear ultimately as lake outflow, include storage as snow-

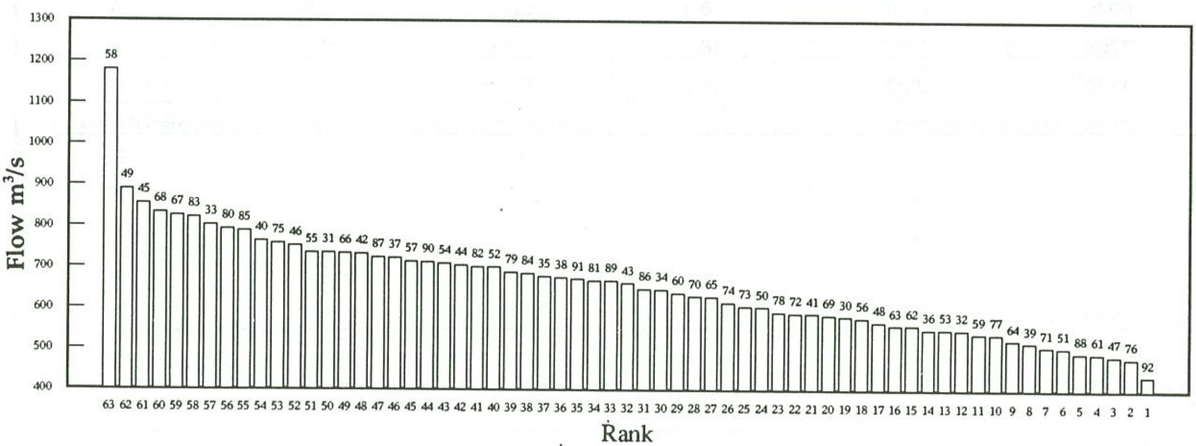

Figure 21.5 Inflow Sum for five lakes (Tekapo, Pukaki (Natural), Ohau, Hawea, Wanaka) - November to May (labels are years eg 92 = 7 months ending May 1992).

pack, evaporation, snowmelt, groundwater recharge, and evaporation from the lake surface. All these are affected by such factors as temperature, cloudiness, wind, anticedent moisture conditions, and so forth. To properly understand the causes of the power shortage requires a sophisticated analysis of the hydrological cycle, and availability of good data on the amounts of water moving through and stored in different components of the cycle.

Moreover, the extreme low inflows recorded in 1991-92 have highlighted the fact that extreme flows, so much a feature of engineering design, do actually occur from time to time. It is easy to forget this!

Acknowledgements

The author is grateful to Eileen McSaveney, Ian Jowett, Paul Mosley and Bryce Carter for reviewing the manuscript.

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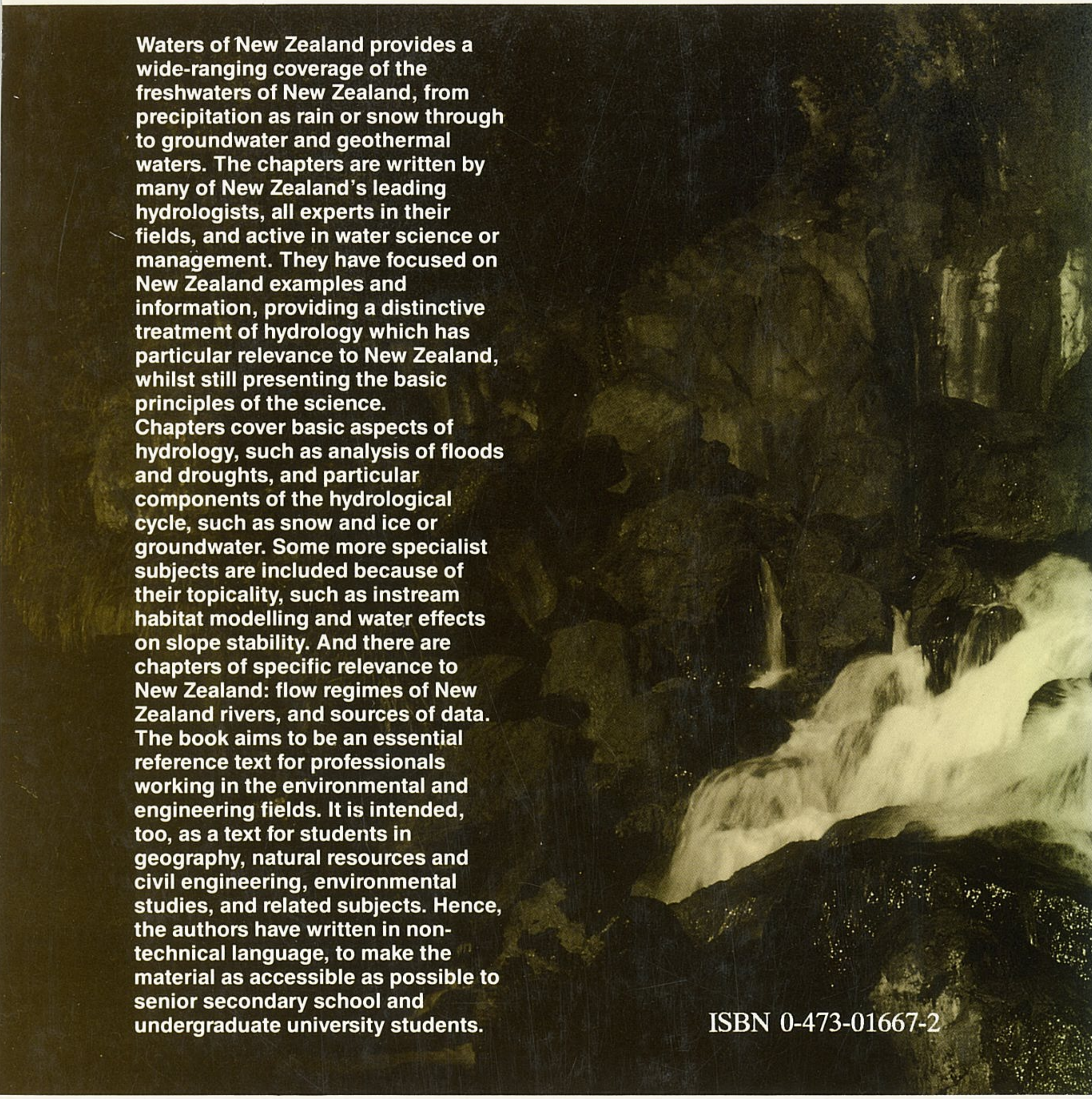
Subject Index

Geographic Index

PRINTED AT THE CAXTON PRESS CHRISTCHURCH NEW ZEALAND



Waters of New Zealand provides a wide-ranging coverage of the freshwaters of New Zealand, from precipitation as rain or snow through to groundwater and geothermal waters. The chapters are written by many of New Zealand's leading hydrologists, all experts in their fields, and active in water science or management. They have focused on New Zealand examples and information, providing a distinctive treatment of hydrology which has particular relevance to New Zealand, whilst still presenting the basic principles of the science.
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ISBN 0-473-01667-2