PHYSICAL HYDROLOGY

New Zealand Experience

Edited by
D. L. Murray & P. Ackroyd

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Toebes Memorial Volume

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PREFACE

Development of hydrology in New Zealand has been largely derivative. In only a few areas have genuine innovations been proposed. This is, of course, logical in the context of our hydrological problems.

This form of development, freely acknowledged by most New Zealand hydrologists, has led to a rather dispersed literature. Despite the derivative nature of progress, we have tackled and solved problems peculiar to New Zealand environments or which have not been explicitly dealt with in the international literature. Publication of these efforts has been spasmodic and has, with few exceptions, merged into the voluminous outflow of hydrological writing. Despite the existence of the Journal of Hydrology (New Zealand) no forum has developed where unashamedly nationalistic reviews, and statements-of-the-art are available.

In 1977 the New Zealand Hydrological Society resolved to support the principle of such a forum. It is fitting that the first results of that resolution should also be a memorial to Cornelis Toebes, the man who contributed so much to the development of hydrological sciences in New Zealand, and who was instrumental in the formation of the New Zealand Hydrological Society.

We wish to acknowledge the support extended to us in this project: the authors were generally enthusiastic and prompt in response to editorial pressure; and the secretarial and cartographic staff of our respective departments responded generously to help overcome our inexperience and technical deficiencies. In particular the Society is grateful to the National Water and Soil Conservation Authority for financial assistance; without support the Society's meagre finances and voluntary staff could not have attempted the venture.

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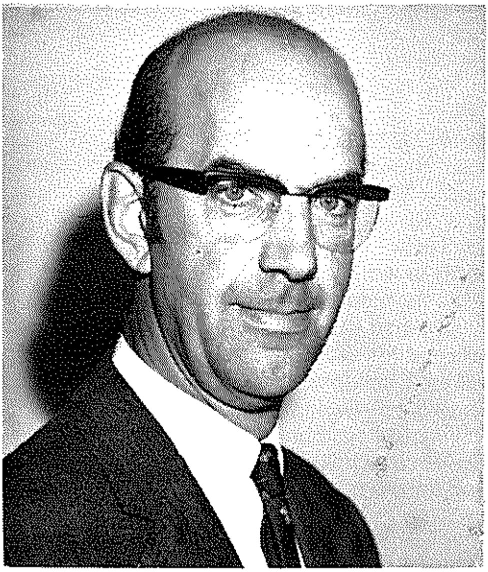
A volume of invited papers in memory of
Cornelis Toebes, 1925–1975,
Foundation President,
N.Z. Hydrological Society

CORNELIS TOEBES, 1925-75



Cornelis Toebes was a remarkable man who made a profound impact on hydrology in New Zealand, Malaysia and Indonesia and who was influential in international hydrological circles. To understand some aspects of his extraordinary attainments it is first necessary to know something of the man.

In the Netherlands

Cornelis Toebes was born at Haarlem, Holland, on 23 June 1925. His father was an architect and he was one of a large family. From 1946 to 1948 he attended the State Agricultural College, Deventer, and graduated with a diploma in tropical agriculture. From June 1948 to July 1951 he was employed by the State Service of Farm Consolidation and Drainage which was headed by Dr W. C. Visser. His work consisted of soil mapping and economic farm research in which the beneficial effect of farm consolidation and soil improvement was ascertained. The soil mapping was done with a number of colleagues of whom Toebes was the chief. In a testimonial Visser stated: "They mapped vast areas according to directions given by Toebes. The work was done with a great degree of independence. Later checks by other soil mapping institutes proved that the work was done with great skill and accuracy, though it took relatively little time." The economic research involved was done by Toebes and a colleague, and Visser recorded that it was "one of the best pieces of work of our office." Toebes received a diploma in engineering in 1950.

In mid 1951 Toebes became restless and declared his desire to explore fresher fields overseas. He decided on New Zealand but had no thought of settling there, his father paying for his passage. During university vacations he had worked on an uncle's farm in the Province of Zeeland. Perhaps this provincial name had stirred his thoughts about New Zealand — a supposed 'south-sea paradise'? However at least seven of his school–time friends came at the same time. And it seems fitting that when he landed in New Zealand he chose the occupation of farm labourer to obtain an entry permit. Toebes arrived at Wellington in late 1951.

At Palmerston North (1952-54)

As a result of high commendation from Dr Visser he was introduced by the Netherlands Consul in Wellington to Mr E. C. Schnackenberg, engineer to the Soil Conservation & Rivers Control Council, which body was serviced by the Ministry of Works. Toebes was employed by the Ministry of Works and posted to Palmerston North to work under the late A. C. 'Hoppy' Hopkins who was Officer–in–Charge of the then North Island Hydraulic Survey. After nearly two years of mainly stream–flow gauging throughout the North Island he left to take up a painting career with a fellow graduate, Piet Vlieg, from Holland. Their business was called the United Painters — echoes of the Netherlands, the United Provinces. Initially they travelled through the streets of Palmerston North in file on bicycles with their painting ladder suspended between them — what a sight! They both qualified as master painters! However this excursion was short–lived because in late 1953 Toebes rejoined Hopkins' team and later was transferred to Whangarei.

At Palmerston North he met Miss Heather Thomson who was a science teacher at the Palmerston North Girls' High School. They married in January 1954.

At Whangarei (1954-60)

On 2 February 1954 Toebes established the Northland Hydraulic Survey of the Ministry of Works at Whangarei, Northland. He quickly organised and implemented a comprehensive programme of hydrological data collection and analysis which has placed Northland in the lead in New Zealand on knowledge of the magnitude and frequency of low flows.

During 1954 he produced his first major New Zealand paper — *Streamflow: polydimensional treatment of variable factors affecting the velocity in alluvial streams and rivers* — which was received on 3 March 1955 by the Institution of Civil Engineers, London. It was published in 1955 and won for him the Manby Premium award of The Institution in 1957. The prize was a high quality slide rule in a leather case — a trophy of which he was justly proud. He had trained mainly in the biological and earth sciences but his knowledge of mathematics was also sound. That after only a few years in a new field of work in a new country he could produce a prize winning paper on the hydraulics of streamflow demonstrated his great adaptability and potential. His chief assistant in the preparation of the paper was his wife, Heather, who transformed the text into acceptable English!

The hydrological requirements for the South Auckland and Waikato regions were undertaken from Whangarei by Toebes. This required him to be away from home frequently for several weeks at a time. Their three children, Justin, Harriet and Quentin, were born at Whangarei.

In 1960 he transferred to the head office of the Ministry of Works at Wellington.

Wellington Years (1960-74)

Toebes subsequently became Chief Scientific Hydrologist of the Water and Soil Division, Ministry of Works and Development (departmental title changed). He was responsible for all hydrological work of the division. To manage this, in later years, he had under his technical control throughout the country a staff of about 33 engineers and scientists and 110 field and data processing technicians. At times his office resembled a railway station as his staff and colleagues sought and went away with his advice. It was during the early years that his ability as a decision maker became evident. Although not everyone liked his decisions, it was these that did so much to establish hydrology in New Zealand as a professional and dynamic field.

In the early 1960s he stimulated national interest in regional hydrology, led a project to demarcate the hydrological regions of New Zealand, launched the representative basin programme and initiated the upgrading and expansion of the experimental basin network. Conscious of the need to train staff he initiated training courses – firstly at Kainga, Christchurch, commencing in 1962 and later in Wellington.

In 1961 the *Handbook of Hydrological Procedures* was commenced and the first procedure, written by Toebes, was distributed. It is noteworthy that of procedure numbers 1–35 which were produced between 1961 and 1965, Toebes was author of 14 and co–author of 8. During these same years he further busied himself to produce reference texts for the Technical Correspondence School, Department of Education, so that in 1963 the two volumes of *Applied Hydrology* were published. Much of this work was done in his own time at home.

On 25 August 1961 the New Zealand Hydrological Society was formed and Toebes was made president. Under his sound leadership a group of eight enthusiasts set about creating a professional society in which standards of scientific excellence were paramount. It was indeed a brave and far sighted move to establish the Journal of Hydrology (NZ) with such a small membership. He was president until late 1973 when he decided to relinquish the position. In the formative years of the Society he gave much – so much that the high status of the young Society, both in New Zealand and overseas, was largely a reflection of the president's great efforts here and of his high personal standing abroad. In 1966 the NZ Hydrological Society became a member body of the Royal Society of New Zealand and in July 1966 Toebes was appointed to represent the Hydrological Society on the Member Bodies' Committee of the Royal Society.

The 1960s was a very busy and challenging period for him but a great deal was achieved and he enjoyed it immensely. He knew that the International Hydrological Decade (IHD) was scheduled to commence at the beginning of 1965 and his aim was for New Zealand to be an active participant. New Zealand did participate in the IHD programme and a national committee was formed which included Toebes as the representative of the New Zealand Hydrological Society. Five national sub–committees were established in 1964 and Toebes was convenor of the one for experimental basins.

His overseas hydrological activities started with his attendance at the IASH Budapest Symposium in 1964. In 1966 he attended the Co–ordinating Council's session in Paris and subsequently represented New Zealand, and at times Australia. Nine international working groups were established for the Decade and Toebes became a prominent member, and later chairman, of the group on representative and experimental basins. This association resulted in the editing by Toebes and Ouryvaev, and the publication in 1970, of an international guide for research and practice, viz, *Representative and Experimental Basins.* His frequent overseas trips during the IHD did much to infuse the New Zealand programme with fresh ideas and also, it has been said, to stimulate the international programme. His international involvement in the hydrological sciences was further expanded when in later years he became an associate editor of the Bulletin of the International Association of Hydrological Sciences (now Hydrological Sciences Bulletin).

The high point of the IHD for Toebes and for hydrology and hydrologists in New Zealand was the IAHS symposium on the results of research on representative and experimental basins held in Wellington during 1–8 December 1970. This had been proposed to the IHD Co–ordinating Council by Toebes and in New Zealand it was organised by the Royal Society assisted by the Hydrological Society. Toebes was chairman of the symposium organising committee and G. Markham of the Royal Society was secretary and chief organiser. The symposium marked the culmination of much planning by many but the prime mover and pusher throughout was Toebes. That symposium was acclaimed widely as a technical success. The pre– and post–symposium tours added further to make the overall venture what some visiting delegates told me was the most rewarding and enjoyable international conference they had attended to date. This was great news for us in New Zealand; it was a wonderful tribute to the leadership and zeal of C. Toebes.

During the mid to late 1960s his visits to Europe and tours to other countries as a member of UNESCO working groups were numerous and on each return to New Zealand he was fired with the latest and best developments he had observed. In New Zealand his knowledge of the international hydrological scene, his enthusiasm and his astute leadership served as a searchlight to probe a long way ahead and to guide hydrology up a basically sound staircase. The upward climb was very rapid – perhaps it was too rapid for adequate consolidation at each step.

As well as the commitments already mentioned this period found him, at some time, deeply involved with various other technical activites which included:

- Chairman, UNESCO international panel on the preparation of a guide on aspects of the influence of Man on the hydrological cycle.
- Member and New Zealand national representative on the international commission on water resources systems and relations.
- Chairman, New Zealand national committee of the International Association of Hydrological Sciences.
- Chairman, New Zealand water quality working group.
- Member, environmental committee of the Royal Society of N.Z.
- Honorary lecturer in irrigation, drainage, hydrology and hydraulics at Lincoln College, University of Canterbury, the Universities of Auckland and Victoria, Wellington, and at Wellington Polytechnic.
- Assessor, in soil and water, for the degree examinations (B.Ag.Sc.) at Lincoln College.

Furthermore, he held a license (No. 5574) as a Netherlands professional engineer and was a member of the American Society of Agricultural Engineers.

From the late 1960s he was in demand by the governments of several southeast Asian countries to advise on water and soil conservation organisation and administration. His reports are testimony of the leading roles he played.

He resigned from the Water and Soil Division, Ministry of Works and Development, in July 1974.

The Final Years (1974-75)

The Wellington office of Tonkin and Taylor, Consulting Engineers, was opened immediately after Toebes left the Water and Soil Division. It was a room in the front of his flat which was later to be shared with two other resigners from Water and Soil Division, viz, Gary Blake and Frank Scarf. Between July 1974 and late 1975 Toebes spent most of his time in Malaysia – his base being the ENEX (The Engineering Export Association of New Zealand) office in Kuala Lumpur. He would return to Wellington for periods of 3 to 6 weeks but after a year his return visits became less frequent because Malaysian demands on his time had increased. Only once did he visit the new office of Tonkin and Taylor in Molesworth House, Wellington.

While in Kuala Lumpur, where he lived in the Federal Hotel, he was in charge of the development of the Malaysian national hydrological organisation and the Kelantan River basin project of which he was the chief planner. Always on the lookout for job opportunities he revisited Indonesia and made a trip to the Middle East to sell water. His second visit to Egypt in December 1975 was to assess the possibility of future work in the Nile Basin, particularly in relation to sedimentation problems. This great river of history must have presented an enormous challenge to him.

Toebes died in Aswan on 10 December 1975. He and three other New Zealanders (Terry Heiler, Vince Bidwell and Max Wigbout) had attended a special reception by local school children in their honour. Toebes returned early to the hotel because of mild dysentery and went to bed. Upon returning about 1½ hours later his room–mate found him dead. He had been writing to his older son, Justin, because it was the eve of his 21st birthday anniversary.

His body was buried in New Zealand.

The Man

I first met Cornelis Toebes in December 1959 in Wellington. We were two of only four hydrologists who gathered with 45 senior engineers and scientists to discuss engineering hydrology problems. It was the first meeting of its kind in New Zealand. For some time it was puzzling to many of us that this tall (1.99 m) Dutchman, with a New Zealand accent, was called 'Case' spelled Kees. That meeting marked his hydrological debut so that in 1960 when, in conjunction with A. P. 'Arch' Campbell, he began to organise hydrology in New Zealand Kees was already a familiar figure to the majority in hydrology at the time.

In those formative years for hydrology he travelled widely and frequently in New Zealand to meet people and see places. He was able to communicate easily with people at all levels – from ministers of the Crown to the most junior staff. As a consequence of this facility coupled with the personal interest he displayed in the welfare of each staff member and his great memory for the personal and family matters of others, Kees gained the high respect of all who worked under his control – and of many beyond it. He was a tenacious task–master but to those who tried hard to do a job well he was generous with praise and encouragement. His thought processes were very much more rapid than those of most, however he was seldom impatient with slower learners.

Of course his memory capacity extended to technical matters. Long afterwards he could narrate step–by–step details of major technical meetings and this was usually flavoured with anecdotes of interest. He possessed an enormous sense of humour which, coupled with his memory for detail, made him a very able and interesting storyteller. Many times at home in the evening, when Kees was visiting the district, my family and I derived much pleasure listening to him tell of his overseas experiences. Among other incidents we heard about the time at a United States airport when the flight he was booked on was about to depart ahead of schedule. Upon his arrival at the airport the plane was already taxiing towards the runway. However he insisted that he must get aboard; so to get rid of one very tall, very forceful and very irate New Zealander the plane was halted. Kees was loaded on a hydraulic forklift, seated on his suitcase, transported to the waiting plane and hoisted aboard! If you can picture this you could also see Kees enjoying it to the full and later laughing uproariously over it.

Kees' deep love of fine art and literature, and of the music of such as Mozart, Beethoven and Bach, no doubt nurtured by his parents in Holland, was matched by his love of the land — the open country, the forested wilderness, the rivers and lakes. His culture extended to linguistic mastery of the Dutch, German, French and English languages, and academic knowledge of Indonesian. Ashley Cunningham recalls an encounter with Kees in Switzerland:

> "During my stay at the Swiss Federal Forest Research Institute I was sometimes invited on field inspections with other visitors to the country. Such an occasion occurred in October 1965, when I accompanied Dr Nageli and US hydrologist Ed Dortignac to visit the Rapengraben study catchment in the Emmental valley. I was told that another New Zealander would join us at Berne; it turned out to be Kees Toebes. Two or three local foresters and hydrologists were also present, and during the morning's field inspection we used English as the common language. Over lunch, however, the party regrouped and conversations developed in several languages. It was typical Swiss hospitality — a lavish luncheon in a specially reserved room in the local inn. Kees was at his most sparkling, talking English to the American and me one moment, and chatting in fluent German to his Swiss neighbour the next. The high point came when Toebes discovered that the young waitress was a student from Holland, whereon he engaged in chatty discussion with her in Dutch, without losing track of his other conversations. This greatly impressed the Swiss, who pride their ability to cope with English, French or Italian, but to whom Dutch is quite a foreign language.
>
> Back at the Institute some days later, I heard Dr Nageli telling his colleagues with obvious respect, of the tall 'Neuseelander' who was able to outdo the Swiss in language, and on their home territory!"

Among his other achievements can be recorded the title, given to him by many of his friends, of 'the world's worst driver' — and for good reasons! As well, he was probably one of the major individual consumers of peppermints in New Zealand, if not the world, and they were carried on all occasions.

In the early to mid 1960s Toebes was seen as a happy man overflowing with 'la joie de vivre'. But as the years went by he seemed to grow more aloof from people and signs of loneliness frequently appeared. Probably this was related to separation from his wife in late 1969 and divorce in late 1974. Perhaps it was also related to administrative complications and frustrations at work. It was a time of great striving and turmoil for him — it was the period in his life when clear identification of priority objectives was needed. There is no doubt that the family separation set him back markedly. To find more purpose in his life he purchased, in 1973, a small farm unit (3.25 ha) called Sedgemoor at Manakau about 75 km northeast of Wellington. This included a historic house built in the 1870s. In weekends he set about to restore the house, reinstate the patches of indigenous vegetation and landscape the pastoral area. However after he started work with Tonkin and Taylor in July 1974 he spent little time on this venture because of his overseas commitments.

Although there is no doubt that he more than amply fulfilled whatever was required of him it seems that by late 1975 he was yearning for greater involvement again in the international hydrological scene. Certainly he was closely connected with the transition from the IHD to the IHP (International Hydrological Programme) but not long before he died he was planning for the rewriting of the UNESCO guide book on representative and experimental basins which he had co–edited. He need not have considered this – there would be no material gain from it – but he saw that it needed to be done and he was responding to that implicit challenge. This was characteristic of him. He determined his own destiny; he was his own chauffeur. But he drove himself too hard for too long. He did it because that was his way; and he did most things very well.

In 1951 Dr W. C. Visser wrote of him: "We appreciated Toebes as a practical man with the ability to do valuable work in scientific research. Toebes has no difficulties in establishing good relations with colleagues; when doing teamwork he has an easy and friendly manner checking working people placed under him. Toebes will show himself in every job demanding intelligence and resourcefulness a valuable man."

In 1978 those who really knew Kees Toebes would acknowledge the truth of Dr Visser's statement. But it should be said that although Toebes had the ability to do research he made his mark in New Zealand in the field of scientific administration. Those associated with hydrology in New Zealand should be thankful that in 1951 Cornelis Toebes left his family and friends in Holland and came here to work for a short time.

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RAINFALL IN NEW ZEALAND

S. M. Hurnard and J. D. Coulter

ABSTRACT

Consistently accurate and representative measurements of rainfall are vital to all hydrological research and application. This paper traces the history of rainfall measurements in New Zealand and outlines some of the problems that have been faced by scientists and technicians working in this field. The major rain-producing processes which affect the New Zealand area are outlined and some characteristics of the intensity, distribution and long term trends in rainfall are discussed.

RAINGAUGE NETWORK DEVELOPMENT

Only unofficial and irregular rainfall measurements were made in New Zealand prior to 1862. In that year regular, official, daily rainfall observations commenced when the first five of a planned network of ten Government meteorological observatories were established at Auckland, New Plymouth, Nelson, Wellington and Dunedin. These were followed in 1866-67 by seven more, while in 1879 the first rainfall stations, known as third class stations, were established. By 1880, there were 14 rainfall stations and in spite of setbacks in periods of economic depression this number rose to 58 by 1891 and to 206 in 1907 when four additional gauges were added in North Otago following a severe drought and an unsuccessful attempt at rain-making by cannon fire. Restrictions brought about by the first World War caused a number of stations to close and further difficulties arose during the brief period of prosperity immediately after the war when rapid changes in land ownership caused many site changes.

The raingauge network was by then well established, however, with sufficient records to compile useful statistics (Robertson, 1950). Demand for these increased with the planning of large hydro-electric power schemes and the movement towards closer settlement and rapid urban development. At their 1918 annual conference civil engineers passed a resolution urging the Meteorological Office to arrange for the collection and analysis of adequate rainfall data. This was in fact being undertaken and by 1927 the number of rainfall stations had risen to 427.

In 1927, the Meteorological Office became part of the newly formed Department of Scientific and Industrial Research (DSIR). This move was followed by a period of very rapid growth, especially in weather forecasting as a result of the increased importance of civil aviation. Climatological data also became increasingly important as agriculture developed towards more intensive farming, better quality products, new crops and improved pest and disease control. Rainfall statistics in particular took on new importance to engineers involved with problems of drainage, flood protection, water supply and stream and river control.

During the late 1930's, an additional 50 rainfall stations were set up by the Public Works Department to obtain information for irrigation and aerodrome construction. Other organisations, for example the Auckland City Waterworks Department, also set up raingauge networks and in most cases records were made available to the Meteorological Office.

The setting up of the Soil Conservation and Rivers Control Council (SCRCC) in 1941 had a significant effect on the growth in the number of rainfall stations. Part of the Council's task was to coordinate and publish existing hydrological data and to recommend improvements to the network of hydrological stations. The Meteorological Service agreed to add a further 400 manual (daily) gauges to those already in existence, to set up 100 additional automatic stations and to coordinate rainfall records from all sources. During this period, a campaign to trace private raingauges throughout the country was undertaken and some further 150 useful records were added to the Service's files. In 1948 new automatic gauges were installed at Paeroa, Rotorua and Waerenga–o–kuri.

Networks of raingauges were set up at the Soil Conservation stations at Makara in 1955 and at Moutere Hills in 1960. These stations were at first operated by the Department of Agriculture and later by the Water and Soil Division of the Ministry of Works. Other gauges or networks were established at Soil Conservation stations and at Experimental and Representative Basins. These have been operated by several agencies (Water and Soil Division, Ministry of Works and Development (MWD), DSIR, Department of Agriculture, N.Z. Forest Service). SCRCC also promoted the measurement of rainfall by Catchment Authorities in previously ungauged high country areas by means of storage gauges.

During the 1950's an effort was made by the Meteorological Service to replace non–standard gauges and to improve and standardise exposures. Today most gauges meet these requirements.

There are still a number of 'observatories' staffed by the Meteorological Service, but the great majority of rainfall measurements are made by co–operative observers, many of them farmers with others employed by government agencies or local bodies, using equipment supplied by the Service.

The Meteorological Service currently collects rainfall data from more than 1500 gauges throughout the country, comprising 'manual' gauges read daily, octapent storage gauges and automatic rainfall recorders. A summary of the types of gauge and the authorities responsible for them is given in Table 1.

At present there are also nearly 900 gauges operated under the National Water and Soil Conservation Organisation. Data from these are held by MWD in the 'TIDEDA' computer files or by local water authorities.

TABLE 1 — Summary of rain gauges in service at 1 January 1978.
(as listed in N.Z. Meteorological Service Register of Stations)

Authority	*Manual*	*Plastic*	*Auto. daily*	*Auto. weekly*	*Auto. long period*	*Octapent*	*Octa auto.*	*Total*
NZ Met. Service	1099	17	131	24	6	104	3	1384
Private	43	24	1	3	3	1	3	78
Private (yearly)	26	—	—	—	—	—	—	26
Catchment	20	1	2	2	3	49	2	79
MWD	249	43	27	55	57	35	430	896

DATA PROCESSING, ARCHIVING AND PUBLICATION

In the Meteorological Service, machine methods of rainfall data processing commenced in 1961, with the punching of monthly rainfall totals. Punching of all daily totals started in 1967 and back punching of key stations has been progressively carried out. Today monthly totals from more than 2600 stations are held on the Meteorological Services magnetic tape files for the total length of station records, as are the daily totals for all current stations (since 1967) and for all climatological stations and selected rainfall stations since the start of records. In total this comprises more than a quarter of a million months of daily rainfall measurements.

Processing is carried out at the Government Computer Centre, currently on the ICL 2980 machine. Data returns are for the most part sent every month by mail and it takes most of the next month to complete the punching. Some rainfall information is received each day as part of the weather reporting organised in support of forecasting. A meteorological computer is now being installed at Kelburn and this is expected to enable a more rapid flow of rainfall and other data for immediate analyses and application.

Data from some recorder charts have been tabulated as hourly rainfall values, but for the most part routine data extraction is limited to maximum falls for each month in periods ranging from 10 minutes to 72 hours. These data form the basis for statistical estimates of depth-duration-frequency relationships (i.e. the rainfall depths for each interval with return periods of 5, 20, 50 years etc).

Rainfall statistics are published annually by the Meteorological Service (N.Z. Meteorological Service, Annual Observations). These include monthly totals from manual raingauges, totals from octapent and storage gauges and maximum rainfall in selected time intervals obtained from automatic gauges. Daily data are not published but may be made available from computer file by arrangement. Summaries and analyses of rainfall data have been issued from time to time (e.g. Coulter, 1969; N.Z. Meteorological Service, 1973; Robertson, 1963).

RAINGAUGES

It is commonly accepted that the earliest measurements of rainfall were made by Castelli in Italy in 1639 using a single glass cylinder about 5 inches in diameter and 9 inches deep. However in a 'History of Korea' published two centuries earlier, mention was made of more elaborate rain gauges 30 cm in depth and 14 cm in diameter. The standard type of gauge now in use in New Zealand is the result of a gradual development from earlier patterns but the basic principles of the instrument remain unchanged from those devised in the 17th Century. Much of this development occurred during the 19th Century and it is recorded that in 1859, G. J. Symons took up the task of standardising the measurement of rainfall in the United Kingdom.

In 1891, the International Meteorological Conference at Munich decided that because of the wide differences in climatological conditions around the globe, it was not possible to lay down a uniform height for raingauges and adopted an earlier resolution that " . . . (gauges) . . . should be sufficiently elevated to be out of the influence of drifting snow and of the splashes of drops from the ground . . . ". The United Kingdom had adopted a height of 12 inches and this lead appears to have been followed in New Zealand from an early date.

There are three main types of gauge in use. The manual gauge is normally a standard United Kingdom Meteorological Office MKII gauge made of copper or stainless steel with a 12.7 cm (5 in.) diameter orifice, read daily. (Plastic gauges, diameter 10.2 cm (4 in.) are sometimes used.)

The standard automatic or recording raingauge used by the Meteorological Service is the Dines tilting siphon gauge with a daily chart, clockwork driven, and it is used to measure duration and intensity of rainfall. The 'Lambrecht' natural siphon recording gauge is used mainly in remote areas as it has a weekly or monthly clockwork chart drive.

Tilting bucket type rainfall recorders are suitable for digital recording (and hence for machine data processing) on punched tape or magnetic tape, and for this reason are becoming more widely used. Amongst other rainfall recorders of some interest, though thought to be no longer in use in New Zealand is the 95 day 'Casella' gauge. This was a very large instrument, with weight–driven clockwork spiral chart drive, designed in 1938 by Mr T. G. G. Beck, an irrigation engineer with the Public Works Department. A number of these produced useful records, but they required very skilful care and maintenance and the pricking stylus often failed to record properly, especially in damp climates. Another was a heated precipitation recorder developed by Batcheler (1970) and used above the bushline at Cupola Basin, Nelson Lakes National Park. This was designed so that the heater burned only when temperature was near or below freezing point and rain or snow was falling. Weighing bucket gauges, much used in the United States of America and Europe have not been widely adopted in New Zealand, although they could be expected to be useful where there is much snow.

The third basic type is the storage gauge suitable for use in remote areas where readings can be taken only at irregular intervals of a week or a month or more. The Meteorological Service uses 'Octapent' gauges which can hold either 680 mm (27 in.) or 1370 mm (54 in.) of rainfall. They have a conventional 12.7 cm collecting funnel, are equipped with a rubber frost protector, and have a narrow neck and flange to reduce evaporation from the inner container. Several of these have been installed in conjunction with automatic rainfall recorders.

A low–cost, long period storage gauge was developed by P. J. Grant and used widely by catchment authorities especially in the North Island mountains. This gauge, commonly called the 'Class C' or 'milk-can' gauge will hold several month's precipitation, and is read with a dip stick when opportunity allows. As the gauge is not emptied by casual observers, a lost reading is of little concern. A film of oil is used to prevent excessive evaporation, and the absence of a funnel makes it a better snow catcher than the more conventional gauge (Grant, 1960). Water and Soil Division and catchment authorities have experimented with 'stand-pipe' storage gauges in the hope of obtaining better measurements of snowfall. Nevertheless precipitation measurement in remote areas where snow and strong winds are common, remains largely an unsolved problem.

For frost protection the Meteorological Service uses 'Killfrost' paste in Dines rainfall recorder floats, and, in a few instances where power is available, has installed an electric light bulb in the housing. Lambrecht floats are protected when necessary by the use of a rubber membrane as the base of the float. Others have experimented with arrangements utilising solar (Stratford and Costello, 1974) and ground heat (Speight, 1962) for frost protection. The highest Meteorological Service recording raingauge is a dines daily installed at Wharite Peak at an altitude of 914 m. A summary of raingauges currently operating at an altitude of over 1000 m is given in Table 2.

TABLE 2 — Summary of raingauges at an altitude of over 1000 m

Authority	*Total*	*Raingauge Type*	*Total*	*Highest*	*Altitude (m)*
NZ Met. Service	24	Octapent	137	Ruapehu Ski Lift	1753
MWD	112	Manual	3	Backridge, Westland	1855
Catchment	17	Automatic	10	Rocky Gully, Otago	1542
		Radio controlled	2		
		Unspecified	1		
	153		153		

Measurement Errors

There are three main sources of error in rainfall measurements, namely instrumental, observational and those arising from raingauge exposure.

Instrumental errors may arise from leaks, evaporation, out–splash, the moistening of dry metal surfaces, icing and overall gauge design. These errors tend to be additive and to decrease the estimated depth of the rain. Grant (1961) suggests that wetting alone may reduce rainfall amounts by as much as 5%.

The single factor having the most serious effect upon the accuracy of the catch is raingauge exposure. Little difficulty arises when rain or snow falls vertically but under windy conditions problems arise. Most experimenters agree that raingauge shields cannot overcome the effects of bad gauge exposure but they may ameliorate them somewhat. Shields have not been widely used in New Zealand.

In a report on investigations into rainfall measurements in the Kaweka range, Grant (1965) considered that of 17 storage gauges, 4 were undercatching by an average of 20%. Similar results were obtained from gauges in the Ruahine range.

Basin or catchment mean rainfall depths for any period may be derived from point rainfalls by, among others, arithmetic mean, isohyetal or polygon methods. Regression analyses relating point rainfall to altitude and exposure have been attempted but in New Zealand where frontal and cyclonic storms have a large convective component, the degree of influence of land relief may be quite variable. Convective storms may be very local, with the cell of heaviest rain sharply demarcated.

Recent Advances in Rainfall Measurement

Two notable recent advances have been the introduction of surveillance radar and the development of automatic weather stations. The Meteorological Service has used radar extensively since the war for measuring the upper wind flow over the country and in 1964 acquired five Cossor CR 353 sets primarily for this purpose. In the following year a surveillance facility was added to detect rain and assess rainfall rates but they were not designed for obtaining quantitative rainfall measurements. Improved radar equipment is now available overseas with a capability of measuring rainfall amounts over a radius of about 100 km with an accuracy to 15%.

The future use of such equipment in New Zealand for short term forecasting of rainfall in cities and for rainfall measurements over certain critical catchments is under consideration.

During 1976–1977 the first ten of some forty planned automatic Weather Stations were installed by the Meteorological Service. These stations are a low–powered meteorological data acquisition system with a range of sensors including an OTA tipping bucket (0.2 mm) rain gauge. Data are at present collected by interrogation over normal telex land lines. To provide for measurements at remote sites a magnetic tape data logging facility is being developed. The feasibility of interrogation by radio either directly or via meteorological satellite is also being investigated.

Estimating Rainfall from Runoff

The mountain country of New Zealand is known to be an area of high precipitation and raingauges operating in the Southern Alps have measured annual falls exceeding 8000 mm. In steep and irregular terrain it is difficult to determine how rainfall varies with height and between ranges. The movement of rain and snow by the wind and their interception by sloping surfaces are additional complications. These difficulties would make the theoretical computations of rainfall depth impossible and point measurements are often unreliable and unrepresentative. However useful information can be obtained from river flow data as recently shown by Jowett and Thompson (1977) for the Clutha catchment. When allowance was made for evaporation, the mean rainfall over the catchment was obtained from measurements of runoff. Raingauges at low levels and in some of the mountain valleys adequately defined the isohyets over much of the catchment, and over the remainder the runoff was found to constrain the isohyets if a plausible distribution was to be obtained. In this way the rainfall on the ranges, east of the main divide was found to be less than had been estimated previously. In fact, east of the divide, the precipitation over the ranges appeared to be only slightly greater than in nearby valleys.

A project to study the processes controlling the distribution of rain over the Southern Alps is being undertaken by the Research and Survey Group of the Water and Soil Division, MWD (McSaveney, 1978; Ministry of Works and Development, 1978). Rainfall is being sampled by 40 storage and 11 automatic rain gauges in a transect across the Alps from Mt Hutt to Waitaha River. From 3000 mm at the western coast, the study shows that the annual rainfall rises to 10,000 mm within 25 km with a narrow peak of precipitation lying about 10 km west of the Main Divide. An equally dramatic decrease in rainfall appears east of this peak; within 20 km of the Main Divide rainfall has dropped to around 1500 mm. Further east again, the study shows that annual rainfall apparently lies within 300 mm of 1200 mm. It was suggested that distance from the western coast may be the dominant control on the location of the high rainfall zone while in the dry eastern high country, altitude and proximity to the eastern foothills are most significant.

RAINFALL MEASUREMENTS IN SMALL CATCHMENTS

Rainfall measurement in small catchments is often difficult because of growth of vegetation and steep, broken terrain may pose problems in the siting of raingauges.

These and other problems have been investigated at the Taita Experimental Station on the eastern hills of the Hutt Valley, about 24 km northeast of Wellington. The experimental area consists of several sub–catchments having a total area of 90 ha at an altitude of between 30 and 230 m above sea level. Many of the slopes exceed 30° and most of the area is covered in scrub or forest. The studies were primarily undertaken to investigate the water and mineral cycles and hydrology of the catchment but were broadened to include comparisons of a number of non–standard gauges and exposures and also the distribution of rainfall under catchment vegetation (Aldridge 1967).

Catches obtained from vertical raingauges were compared with those obtained from directional gauges (vectopluviometers), tilted gauges (orifice parallel to the slope) and ground level gauges. The variation of catch within the network of vertical gauges was small and for most purposes an adequate estimate of the mean catchment fall could be obtained from the rain gauge in the nearby climatological enclosure.

Vectopluviometers made it possible to determine the patterns of direction and inclination of the rainfall and measurements so obtained were used to characterise storms and predict slope rainfalls. Aspect distribution correlations with wind run, and seasonal patterns were thus found. Because rainfall direction within the catchment was strongly modified by the local orography, vectopluviometers did not satisfactorily predict differences between individually paired vertical and tilted raingauges (Aldridge, 1975).

In the studies at Taita, an overall difference over a flat grassed surface between

0.3 m and ground level gauges was found to be 6.3%. This agrees with the findings of other workers both in New Zealand and overseas, but direct comparisons with other studies is difficult because of differences in raingauge design and exposure practices. Most such studies show deficiency in catch of raingauges exposed above the ground in relation to those exposed at ground level, especially when wind accompanies the rain. A difference of 13.5% was obtained between gauges at 0.3 m and ground level on a steep hill site, whereas the difference between the tilted 0.3 m level gauge and ground level gauge was only 2.9%. It is not possible however to state whether this smaller difference in catch was typical of all hill sites (Aldridge, 1976).

In order to investigate fully the water and mineral cycles and hydrology of a catchment it is necessary not only to measure the rainfall over small catchments but also to gain information on the distribution of that rainfall on to the soil surface beneath any vegetation cover. The two small catchments concerned in this study had an almost complete cover of scrub and forest and rainfall interception measurements were carried out on manuka, gorse, hard beech, kamahi and on the vegetation of an exotic catchment. Rainfall was measured above and below the canopy with the contribution of through–fall and stem–flow measured independently. The studies have shown that stem–flow is a significant factor in the distribution of water to the soil (Aldridge and Jackson, 1973).

RAIN PRODUCING PROCESSES

Any rain producing process must include a mechanism for obtaining upward motion of the air. The major mechanisms are extra–tropical cyclones and fronts, tropical cyclones, convective activity in unstable air masses and orographic lifting.

In the New Zealand area, extra–tropical cyclones or depressions and their associated fronts are major rain producing systems. In these, a broad scale ascent of air takes place in areas of strong low–level convergence accompanying transient waves in the upper level westerly flow. Such conditions usually develop most strongly beneath the zone of a maximum wind speed in the upper westerlies (the 'jet stream'). Here baroclinic instability due to strong horizontal temperature gradients leads to a rapid increase in the amplitude of the upper level wave with a corresponding increase in cyclonic circulation around a deepening low pressure centre at the surface.

These large scale cyclonic systems also produce the fronts which in their turn may give birth to smaller wave depressions. Modern meteorological theory oriented to computer mathematical modelling has tended to discount the significance of fronts in favour of the dynamics of large scale motions. Nevertheless, although our knowledge and understanding of fronts remains incomplete, they remain a basic tool of the weather analyst. They mark surface wind and pressure discontinuities, define the detailed structure of rain bands and usually delineate a change in the character of the weather. 'Warm' fronts and 'occlusions' appear less frequently on southern hemisphere weather maps than in some parts of the northern hemisphere. This is in part because of the more zonal nature of the wind flow and the smaller oceanic–continental contrasts of the southern hemisphere.

New Zealand lies near the southernmost latitudes reached by tropical cyclones before they are transformed into extra–tropical depressions. Although as many as seven or eight storms may develop over the southwest Pacific in a typical hurricane season, only one or perhaps two will directly affect New Zealand each year. However they make a significant contribution to the rainfall and its relatively high variability in summer in northern districts.

Convective activity in unstable air mass conditions also contributes significantly to the rainfall in many parts of the country. It was recently estimated that non–frontal convective activity accounted for from 14 to 20% of the annual rainfall over the three North Island hydro–catchments of Waikato, Taupo and Rangitaiki (Hurnard, in prep.). This was similar to the estimated percentage of annual rainfall contributed by depressions of both tropical and extra–tropical origin. This study, covering the period from 1968 to 1975 also indicated the importance of orographic lifting as a rain producing process. Northerly, pre–frontal, orographically enhanced rain accounted for 12 to 18% of the average annual catchment rainfall and in some years this figure was as high as 25 to 30% for the Rangitaiki catchment. Orographic lifting is probably of equal significance in the east of the North Island and of even greater significance as a rain enhancer in western areas, particularly of the South Island.

The distribution and intensity of frontal rain is also very much dependant on orography. For example, in the study previously quoted, it was found that in the period 1968 to 1975 fronts moving in a westerly or northwesterly air flow produced an average fall of 12 mm of rain over the Taupo catchment but a mean fall of only 6 mm in the Waikaremoana catchment. Conversely, fronts which were accompanied by a southerly wind change along the east coast of the North Island produced on average 8 mm of rain over the Waikaremoana catchment while a mean fall of only 2 mm was recorded over the Waikato and Taupo catchments.

RAINFALL CHARACTERISTICS

Milford Sound is the wettest place in New Zealand for which long term rainfall records are available. On April 17th 1939, 560 mm of rain fell in 24 hours and on 13 February 1958, 240 mm fell in 6 hours and 354 mm in a 12–hour period. This of course occurred in an area of high annual rainfall and in general it can be said that the most intense daily rainfalls occur in the wettest parts of the country. However areas of lower annual average rainfall also experience intense daily falls; such areas are to be found in northern Hawkes Bay, and about the Coromandel Peninsula. By contrast, daily falls reaching 80 mm are rare in the Manawatu, Otago and Southland. Analyses of rainfall depth–duration–frequency relations can be prepared for most of the recording rain gauge stations shown in Fig. 1, covering rainfall duration from 10 minutes to 72 hours and for return periods of from 2 to 50 years (Robertson, 1961).

The average duration of rain may also be assessed and it has been established that on average, there are only small differences in diurnal variation in rainfall. Northern areas generally have a maximum during the afternoons as the result of convective activity while districts around Cook Strait and in the west and south

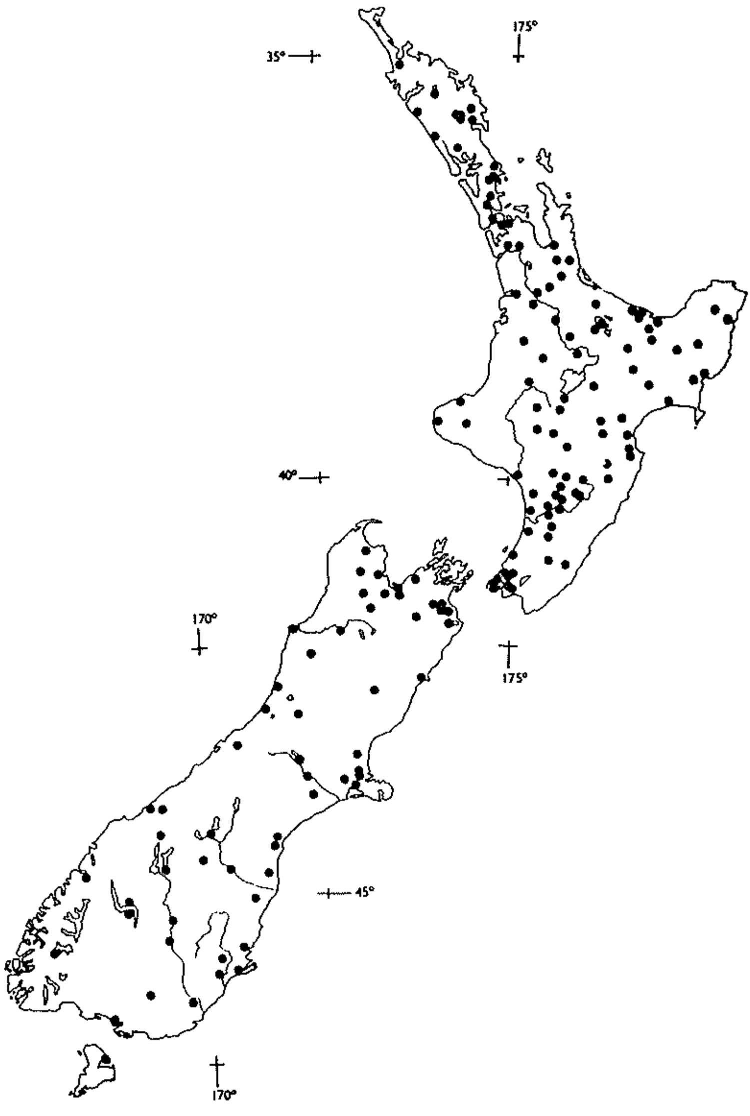

FIG. 1. Recording raingauges as listed in the N.Z. Meteorological Service register of stations, November 1977.

of the South Island experience 54–57% of their rainfall at night in spring and summer.

Seasonal Rainfall

On average, rainfall is spread fairly evenly over the year. Winter rain predominates in most areas and in the north twice as much rain falls in winter as in summer. On the other hand, winter is the season of least rainfall over much of the southern half of the South Island where a summer maximum is found inland, partly the result of convective showers. Spring rainfall is generally enhanced in the west and diminished in the east corresponding to a seasonal maximum in the westerlies during October and November (Fig. 2).

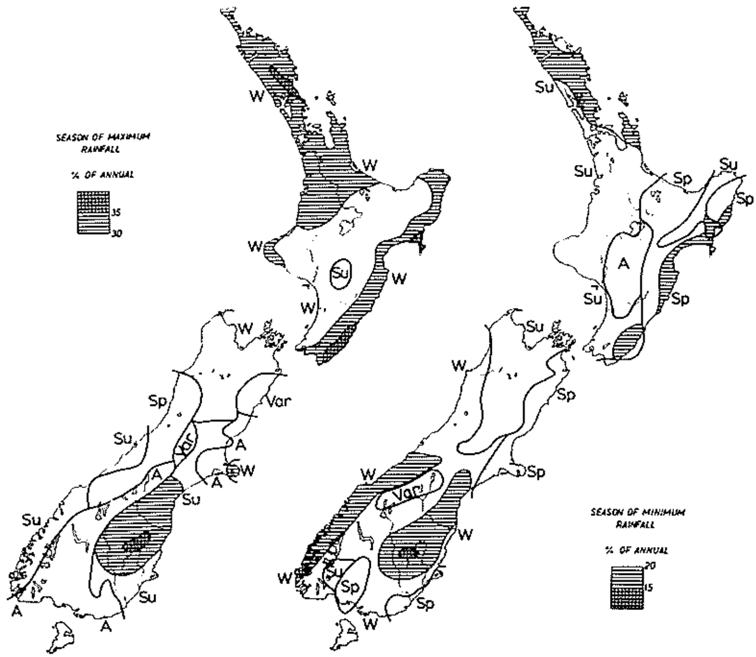

FIG.2— Seasonality of rainfall (based on 1941–1970 data).

In general, seasonal and monthly rainfall distributions are skew and can be represented by a 'gamma' frequency curve. The coefficient of variation of seasonal totals ranges from 0.2 in winter and spring in western areas to about 0.5 in summer in northern and eastern North Island districts. Similarly for monthly totals, coefficients of variation of 0.8 to 0.9 have been found for stations in the north and east of the North Island in late summer and between 0.3 and 0.5 in western and southern districts in the winter and spring. In the North Island, variability tends to be greatest in summer and least in winter whilst in the South Island it is greatest in summer and autumn and least in spring (Coulter, 1969).

A slight persistence in monthly rainfall is apparent such that for a typical station, the probability of any month having below average rainfall is about 0.55 if the preceding month was 'wet' and 0.60 if it was 'dry'. Runs of 'dry' months are more common than runs of 'wet' months.

Annual Rainfall

The mean annual rainfall over New Zealand ranges from 350 mm in Central Otago to over 8000 mm in the Southern Alps. Most of the country receives 800 to 1600 mm of rain on average but those areas in the east of both islands receiving less than 800 mm experience frequent dry spells and droughts which necessitate special dryland farming techniques of supplementary irrigation, and summer dry periods are also a significant factor in reducing production in many other farming areas with higher average annual rainfall. Areas with over 3200 mm are for the most part mountainous and unoccupied.

Annual rainfall totals are least variable in the southwest of the country and most variable in the east and north. These totals have an almost normal distribution with coefficient of variation usually falling in the range 0.12 to 0.25. Totals in individual years rarely exceed 175 percent of the station average or fall below 50 percent. These limits have been surpassed only in low rainfall areas from South Canterbury to East Cape.

Motions in the atmosphere occur as waves. The combination of an anticyclone and depression is one example which has a typical wavelength of 2000 to 5000 km. Waves longer than these also exist but are less marked in the Southern Hemisphere than in the northern. Nevertheless they do occur and the pattern into which they fall tends to persist for several months before abruptly changing. These very long waves, of which there may be only three of four around the hemisphere, are not directly related to the day to day weather but act as steering currents that determine storm tracks and the most favourable positions for storms to develop and decay. It is suggested that changes in their amplitude, phase and wavelength cause much of the year to year variation in rainfall, and may account for any long term trends (Trenberth, 1977).

Long term trends for several areas of New Zealand have been studied, based on annual rainfall indices covering the period from 1890 to the present day (de Lisle 1961). These show short term fluctuations to be unrelated between areas but for periods greater than 12 years a relationship does exist in northern areas probably due to variations in the number of tropical depressions passing across northern New Zealand. There is little evidence of any long term trend.

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SNOW HYDROLOGY

B. B. Fitzharris

ABSTRACT

Snow hydrology is in its infancy in New Zealand with few studies and measurements. Information on snow is seldom used in water management decisions. This situation has arisen because of lack of agreement on the size and significance of the snow resource. To help resolve this issue the literature on snow and ice studies in New Zealand is reviewed. The water equivalent of perennial snow and ice has been estimated as 50 km^3, the bulk of it in the Waitaki catchment. Assessing the seasonal snow resource is more difficult, but after reviewing general reports, water balance models, and analyses of runoff records, it is concluded that for South Island mountain catchments, snow storage supplies 10–25% of annual streamflow. Little is known of the more important year to year fluctuations in the seasonal snow resource. Implications for operation of hydro-electricity reservoirs are briefly discussed.

Methods for measuring the snow resource are reviewed, and assessed for New Zealand conditions. Those include index and sample snow courses, continuous remote measurement, aircraft and satellite observations. Modelling is a powerful potential tool for continuous monitoring of the seasonal snow cover. Important questions for the future include possibility of snow pack augmentation with orographic seeding, management of the snowpack, the effect of land use change in the seasonal snow zone on water yield, and the influence of snow structure on hydrologic response.

INTRODUCTION

In many Northern Hemisphere countries, where the role of seasonal snow melt in flood hydrology is well defined, much effort is invested in measuring the accumulation and rate of melt of the snowpack. An extensive series of snow courses is monitored each month during winter and spring in the Pacific North West of North America. These data are used to aid flood forecasting, to ensure optimum management of storage reservoirs, and to allocate water among competing users. Similarly, extensive routine measurements of seasonal snow cover and of the mass balances of glaciers are accepted responsibilities of authorities responsible for hydro-electricity generation in Norway and Switzerland. Much research effort is devoted to modelling the build up and melt of the snow cover, and vigorous attempts are made to improve techniques of assessing the size and timing of the annual snow melt.

In New Zealand, the state of snow hydrology is still in its infancy, with few regular measurements and very limited attempts to model the seasonal.snow cover. Studies have been made of the mass balance of the Whakapapanui Glacier, Mt Ruapehu (Kells and Thompson 1970). The Ministry of Works and Development has monitored the mass balance of the Tasman and Ivory Glaciers (Anderton 1975, 1976a, b; Anderton and Chinn, 1978), and with the Forest Service maintains various snow courses in the Fraser, Waitaki and Waimakariri catchments, but the information does not appear to be used in decision making by water users on a routine basis.

Reasons for Limited Interest

This contrast between New Zealand and overseas countries appears to have arisen for two reasons. First, we cannot agree in New Zealand as to the size and importance of the seasonal snow cover. The New Zealand Country Report on

Water Resources (1977), our official statement on water, recognises that lake storage is important, but says that "seasonal snow is very variable and of little significance". On the other hand, during the dry weather over the South Island hydro storage lakes in spring 1977, the Otago Daily Times reported on the 12 September 1977, that

> "the Minister of Electricity, the Hon. G. F. Gair, said he understood it was necessary to lower lake levels slightly to meet demand until either rain fell in the South Island or the spring thaw started."

Clearly runoff from seasonal snow cover was of considerable significance in this context.

Second, as Gillies (1964) noted in his study in Central Otago, snow accumulation "is not a continuous cumulative process as in continental snowfields; but it thaws, recedes and reforms at intervals through the winter, so that the water content of the snowpack is variable and hence difficult to measure." The maritime location of New Zealand mountains means the snowpack accumulates and melts erratically. Hence there is no pronounced annual flood in our rivers directly attributable to snow melt, as there is in the Northern Hemisphere countries where the major flows are associated with the spring runoff. Where such snow melt floods do occur in New Zealand they are often associated with heavy, northwest rains along the main divide, so are less obvious, and are considerably damped by storage provided by large natural lakes.

The New Zealand National Committee for the IHD (1969) also noted that snow surveys are of most value where there are heavy winter snowfalls and hot dry summers. New Zealand's maritime climate, with rainfall well distributed throughout the year so that river flows are never dependent on snow melt alone, precludes the direct application of techniques developed in continental parts of North America. Snow melt and deposition both occur at any time throughout the year, so it is impossible to recognise exclusive accumulation and melt seasons. Except in Central Otago, the principal areas of snow tend to be rugged, so that snow cover is highly variable over space, and difficult of access. There is little or no tall vegetation in New Zealand seasonal snow regions, and winds tend to be persistently strong, so that there is considerable snow movement and loss. Finally, good flow gauging sites in the rivers which drain most alpine areas in New Zealand are rare. Consequently accurate flow records from snow catchments are difficult to obtain.

NEW ZEALAND SNOW STUDIES

Snow regions

The New Zealand National Committee for the IHD prepared a tentative classification of snow regions in 1969 (Fig. 1). Owing to the paucity of information on the distribution of snow and its variability in time and space, the classification is broadly defined using arbitrary criteria:

A. Permanent snow and ice. Glaciers and mountain areas where snow or ice persists throughout summer and autumn in most years.

B. Seasonal snow cover. Areas where snow persists for an appreciable period in spring in most years. The lower boundary was suggested at about 1000 m in the south of the South Island and about 1400 m in the North Island.

C. Occasional snow cover. These areas extend below the seasonal snow cover, reaching sea level in the south and east of the South Island. This snow may be significant in some years, especially with regard to stock losses and local flooding, but generally is of little importance for runoff.

D. Snow free areas. Where snow never or rarely lies.

The classification is largely based on similarity in meteorological exposure, the observed pattern of snow cover, terrain, and altitude. Snow regions were further subdivided into subregions, although the boundaries of these were expected to change as more data became available. Snow information has been published for a number of these regions (Table 1).

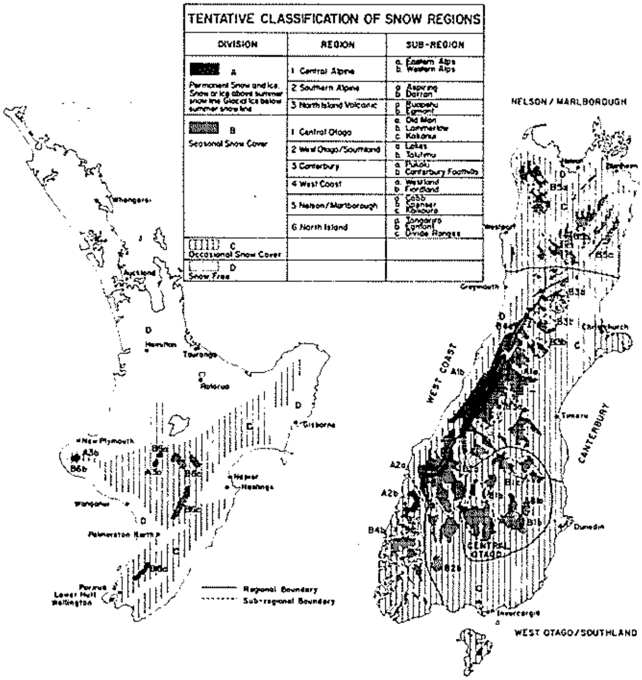

FIG.—1 Tentative classification of snow regions. After New Zealand National Committee for the IHD (1969).

TABLE—1 Published information for snow regions of New Zealand.

Region	*Author*
A. *Permanent Snow & Ice*	
1. Central Alpine	Anderton (1973) (1975) (1976a, b), Anderton & Chinn (1973, 1978), Chinn and Bellamy (1970), Dickson (1974), Goldthwait and McKellar (1962), Sara (1968, 1970), Skinner (1964), Harding (1972)
2. Southern Alpine	Bishop (1977)
3. North Island Volcanic	Heine (1963), Kells and Thompson (1970), Krenek (1959), Odell (1955).
B. *Seasonal Snow Cover*	
1. Central Otago	Fitzharris (1976), Gillies (1964), Jowett and Thompson (1977)
2. West Otago/Southland	
3. Canterbury	Anderton (1974), Archer (1970), Chinn (1969), Morris and O'Loughlin (1965)
4. West Coast	
5. Nelson/Marlborough	
6. North Island	Heine (1962).
C. *Occasional Snow Cover*	Burrows (1976), Chinn (1968), Hughes (1969) (1974), Neale and Thompson (1977), Tomlinson (1967).

Snow and Streamflow

Snow information is still lacking for large areas of the country, some of which, such as the catchments of the Buller, Clutha Lakes and Waiau, are important hydrologically. Few studies have been able to show whether the different snow regions of Fig. 1 give rise to different hydrological response. Anderton (1973, 1974) and Murray (1974) have pointed out that seasonal patterns of streamflow vary, depending on the relative proportions of permanent or seasonal snow. Two contrasting types of distribution are illustrated by the Hooker River and the Manuherikia River (Fig. 3). In the former, maximum flows occur in January, when almost 20% of the annual yield passes the gauging station. There is a steady decrease to June, July and August, in each of which approximately 2% of the yield occurs. In winter, most precipitation in the Hooker catchment is as snow, and melt rates are negligible. Streamflow through the year therefore closely follows the insolation cycle.

By contrast, in the Manuherikia catchment, where there is no area of permanent snow, but extensive seasonal snow, the runoff pattern is one of seasonal snow melt with some rainfall at lower altitudes in winter. Peak flows tend to occur in October in response to rising temperatures in the spring, and there is a rapid reduction in flow during the summer to a minimum in February. Many South Island mountain catchments fall between these two examples.

The seasonal mean flow patterns in the sub–catchments of the Clutha have been investigated by Jowett and Thompson (1977) and they show that average altitude is an important control (Fig. 2). All rivers have low flows in July, the month with lowest average temperature. After July, the energy available for snow melt increases. The lower catchments, such as the Manuherikia have their highest rate of snow melt and runoff in September, the intermediate catchments such as the Arrow and Fraser a month later in October, while the higher mountain catchments of Wanaka, Wakatipu, Hawea and Shotover have their runoff peak in November.

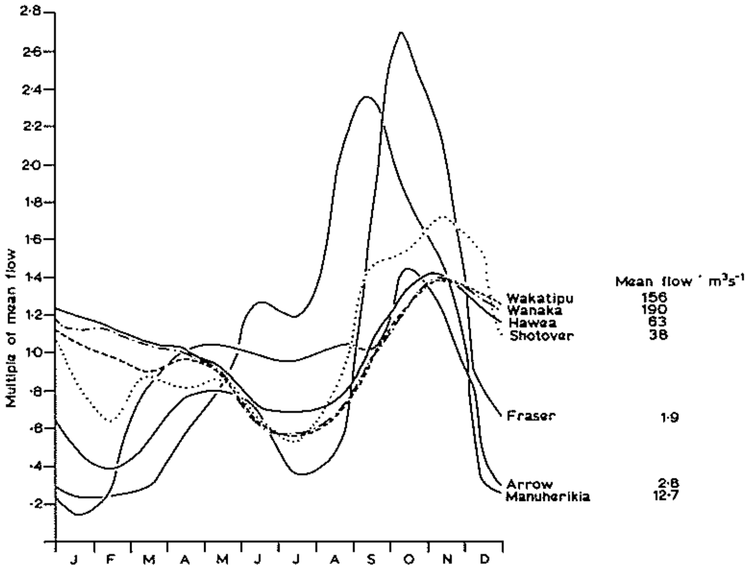

FIG.—2 Comparison of seasonal variation of mean flow in the Clutha River subcatchments. After Jowett and Thompson (1977).

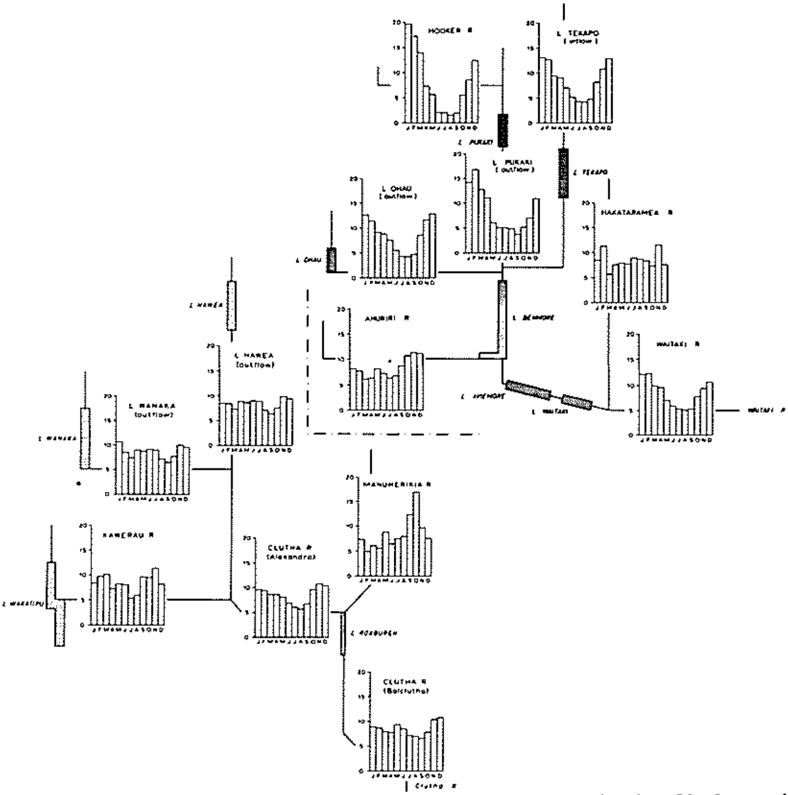

FIG.—3 Mean monthly water yields for thirteen stations in the Clutha and Waitaki Catchments. The seasonal patterns of stream flow for major tributaries and for the principal rivers are shown by expressing mean monthly water yield as a percentage of the mean annual yield at each station. After Murray (1974).

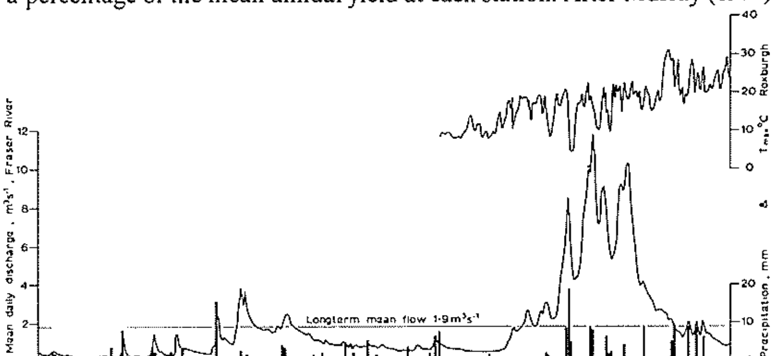

FIG.—4 Typical runoff pattern for a high mountain catchment: daily mean flows for Fraser River for 1974. Daily precipitation and maximum temperatures are shown for comparison for Roxburgh, a nearby low elevation climatological station.

The Fraser River shows the extreme of a high catchment with very pronounced snow melt and freeze. Diurnal fluctuations are evident in the flow record from this river. Plots of daily flows for this, and other South Island rivers with substantial parts of the catchment above 1000 m show an upward bulge in base flow in spring and early summer associated with rising temperatures (Figs. 4 and 7). Superimposed on this upward bulge are a series of spikes, which mark the input of liquid precipitation from storms, plus some unknown component of snow melt associated with energy transfer to the snowpack mainly from condensation and sensible heat. These patterns are repeated annually, but there is considerable variation in actual discharge from year to year.

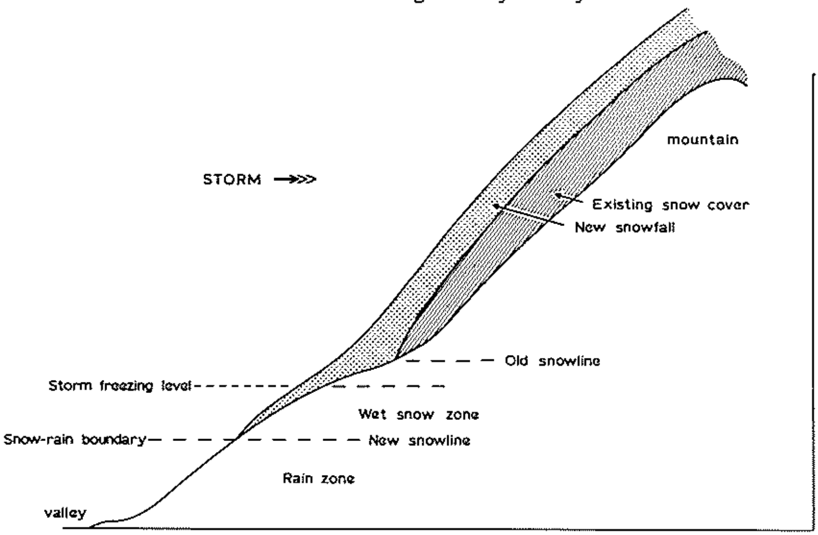

FIG.–5(a) Schematic cross section showing how snow deposition accumulates with elevation on a typical New Zealand mountain after a storm. Effects of wind drifting and aspect are ignored. Snow depths are greatly exaggerated compared with height of mountain.

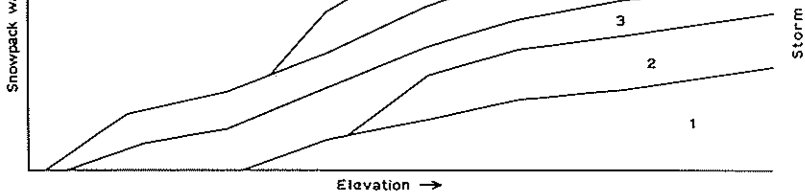

FIG.–5(b) Illustration of how successive snowfalls (numbered (1) to (5)) from storms with different freezing levels combine to produce a snow wedge of increasing snow accumulation with elevation.

Weather and Snow

New Zealand mountains may be broadly defined as west coast and mid–latitude, which implies that freezing levels, and hence the snowlines during most winter storms intersect the mountain at some elevation above their base. As falling snow passes below the freezing level it begins to melt, and finally becomes rain. The net result is that after a snowfall there is usually no snow at the base of the mountain. Higher, there is often a well defined new snowline above which snow depths increase steadily with elevation in the shape of a snow wedge. Where snow has fallen in the melting layer immediately below the freezing level, it is referred to as the wet snow zone. Fig. 5a shows how the lower limit of new snow is closely controlled by the storm freezing level. Each successive storm will tend to have its own snow wedge, of different shape and with different snowline (Fitzharris 1975). Along the Main Divide, storms incorporating a typical Tasman Sea frontal system may produce a compound snow wedge: a higher elevation wedge produced by snow in the prefrontal northwest airstream; then superimposed on this a wedge with lower elevation snowline associated with the colder air behind the front.

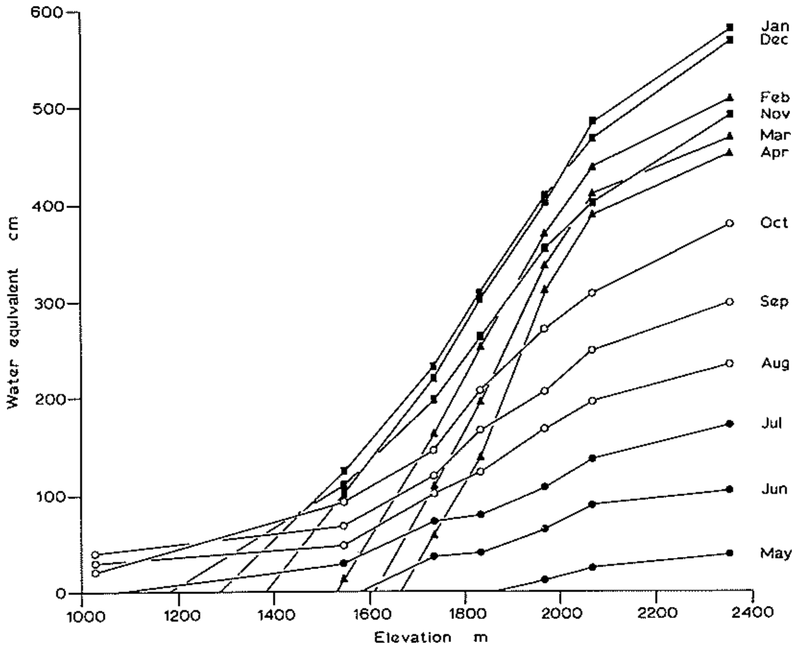

FIG.–6 An example of the variation of snow pack water equivalent with elevation for a New Zealand mountain. Data are for the Tasman Glacier for May 1968 to April 1969 (N.Z. Hydrology Annual) and are adjusted to the first day of each month.

The wedges of new snow from successive storms produce a similar wedge shape increase in snow pack water equivalent with elevation (Fig. 5b). Inter–storm melt effects, which tend to be more pronounced at lower elevations, act to steepen the wedge. The shape of this snow wedge does not behave consistently from month to month because it depends on the nature of the weather (Fitzharris 1972). There is a tendency for the slope of the snow wedge to increase during the first half of the snow season, and to remain constant or slightly decrease during the latter half (Fig. 6), but its exact behaviour differs from year to year.

Three principal factors control the shape of the snow wedge:

(a) Variation of precipitation with elevation. On most mountains there is a steady increase of precipitation with elevation. However the rate of increase, and the elevation of any maximum of precipitation, will change with the stability and wind shear of the storm air masses (Elliott and Shaffer, 1962; Sawyer, 1956). Hence the frequency of various storm types determines the variation with elevation of total winter precipitation, and the snowpack.

(b) Variation of snow melt with elevation. From year to year the energy available for melt is not the same at each elevation.

(c) Elevation of the rain/snow boundary during successive storms. This boundary is largely controlled by storm freezing levels. In New Zealand winters, freezing levels intersecting the mountains may move over a wide range of elevation in the space of a few days (e.g. ahead of and behind a family of east moving fronts). The frequency of winter storm freezing levels at each elevation, together with the magnitude of the storm snowfalls is probably the main factor controlling the shape and slope of the snow wedge.

Local variations due to different aspects, slopes, and snow drift are as important as those due to elevation. Hence the snow wedge concept is best applied to homogenous terrain segments, or in general terms to large catchments.

Because there are few climatological stations above 1000 m in New Zealand, there has been little opportunity to document the weather conditions of snow storms in our mountains. Daily maximum and minimum air temperature, precipitation, and wind run are available at 1400 m for the Ivory Glacier for the 1977 winter (Anderton 1976a). Temperatures fluctuate around 0°C, and rain can occur in midwinter. Conversely snow may fall in summer. The estimated annual precipitation for 1971–75 was 9630 mm, with snow representing 25% (Anderton and Chinn 1978).

Exceptional snow storms in the South Island high country have been documented by Burrows (1976), while detailed analysis of unusual snowfalls at low elevation have been completed for some recent events (e.g. Chinn, 1968; Tomlinson, 1970; Hughes, 1974). Neale and Thompson (1977) give short synoptic histories of the ten heaviest low country snow storms in southern New Zealand in the decade 1966–75. Three conditions are essential for these storms to occur: strong ascending motion; very cold temperatures in the lower troposphere; and an abundant supply of moisture. At least three other factors can influence the distribution, duration and intensity of snowfall: the downward penetration of snow caused by cooling of air in the melting layer; the seeding of a low stratiform cloud bank with ice crystals from an upper cloud layer; and the effects of topography.

Microclimatic studies on the Ivory Glacier (Anderton, 1976a, b; Anderton and Chinn, 1978; Dickson, 1974; Harding, 1972) are the only published information of the energy sources for snow and ice melt in New Zealand, although Kells and Thompson (1970) discuss their influence on the Whakapapanui Glacier. Ivory Glacier studies show that the net radiative flux is dominant, accounting for over 50% of the energy available for melt. Sensible heat flux ranked second in importance (29%), but latent heat flux was also high (17%), and greater than for overseas temperate glaciers, reflecting the maritime environment of the Ivory Glacier. Energy attributable to the heat content of precipitation was not significant for the 47 days studied. However, the heat content of precipitation may be significant during major rainstorms, and in spring and autumn months of high precipitation.

Perennial Snow and Ice Resource

Anderton (1973) estimates the volume of perennial snow and ice in the glaciers of the Southern Alps to be 63±4km^3, most being concentrated in the Waitaki catchment (Table 2). Assuming a density of 850 Kg m^{-3}, the stored water is over 50 km^3. Under present climatic conditions these glaciers are releasing water from long–term storage in excess of the input from snowfall. The quantities of water derived from this depletion of storage make significant contributions to the flow of the Waitaki and certain West Coast rivers. If the estimated average net balance for the Tasman Glacier of about 1 m water loss over the surface area (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 storage in these glaciers is about 1.9 x 10^8 m^3. If this melt water is released from January to March, then the contribution is equivalent to a mean flowof 24 m^{3}s^{-1}, and represents about 10% of the inflows to the lake for this period. The contribution would be greater in dry summers.

Although quantities of water released from perennial snow and ice tend to be less important for other catchments, they can be influential in regulating streamflow. Melt water from this source tends to increase summer flows, particularly during dry summers and decrease winter flows relative to glacier free catchments.

TABLE–2 Glacier resources of the Southern Alps. After Anderton (1973) (Glacier volume is given as volume of ice).

East coast basins			*West coast basins*		
Basin	*Glacier area (km^2)*	*Glacier volume (km^3)*	*Basin*	*Glacier area (km^2)*	*Glacier volume (km^3)*
Waitaki –			Cook –		
Lake Pukaki	186.9	24.58	Fox	41.7	5.27
Lake Tekapo	43.1	2.58	Others	12.9	0.93
Lake Ohau	20.7	0.71			
			Cook total:	54.6	6.20
Waitaki total:	250.7	27.87			
			Waiho	57.8	5.61
Clutha –			Arawata	56.2	3.82
Lake Wakatipu	38.1	2.22	Whataroa	47.7	3.28
Lake Wanaka	32.5	1.32	Karangarua	34.6	1.68
Lake Hawea	4.2	0.12	Wanganui	39.2	2.11
Others	2.9	0.12	Waiatoto	27.1	1.51
Clutha total:	77.7	3.78	Haast –		
			Landsborough	30.4	1.06
Rakaia	43.6	2.26	Hollyford	11.6	0.41
Rangitata	39.4	1.59	Milford Sound	7.7	0.33
Ashburton	4.2	0.20	Paringa	6.1	0.24
Waimakariri	4.3	0.11	Hokitika	5.9	0.18
			Waitaha	4.5	0.13
			Cascade	2.9	0.11
			Turnbull	1.4	0.06
			Mahitahi	1.9	0.04
East coast total:	419.9	35.81	West coast total:	389.6	26.77

Seasonal Snow Resource

Whereas the general size of the perennial snow resource is known, much controversy surrounds the size and significance of the seasonal snow cover.

General reports

Many authors have referred to the influence of seasonal snow on runoff. Gilkison (1930), referring to the great Clutha flood of 1878, comments

> "The winter of 1878 was a long and severe one. Great falls of snow took place on the mountains, and it was said in some ravines the drifts lay 100 ft thick. Hard frosts and further falls consolidated the snowfields, and those who knew Otago well prophesied a flood unless the spring weather should allow it to get away gently and gradually. The conditions which followed were the worst possible. All through the winter and early spring the snow lay on the ranges. In September came a hot nor'west wind which blew for several days and commenced a general melting process; this was followed by thirty-six hours of continuous warm rain, which caused a very rapid thawing of the snow. On the 26th September the Molyneux began to rise ominously and rapidly, and soon that great river was in full flood."

Further rain fell to make the flood the largest known on the Clutha (3600 $m^3\ sec^{-1}$ at Alexandra). Burrows (1976) also noted 1878 as a year of exceptional snowfalls and cold.

Trenberth (1977) has investigated interrelationships among monthly and daily time series of inflows to the Clutha Lakes, broad scale atmospheric indices, and rainfall using spectral and cross–spectral calculations. He estimates that a third of the daily variance of Clutha lakes inflow is unrelated to rainfall or the annual cycle. This unexplained variance is attributed, in part, to the effects of snowfall and accumulation. Lag relationships revealed that rainfall in the catchment area reaches the lakes about a day after it falls, but lead times of "up to ½ month are evident for longer time scales, no doubt caused by the time taken for snow to melt."

Water balance approach

Several attempts have been made to assess the water stored as snow or ice using a water balance approach. The annual water balance is expressed as:

$$\text{input} = \text{output} \pm \text{storage change}$$

$$\text{or} \quad P = Q + E + (\pm G \pm S \pm I) \qquad (1)$$

where P = precipitation
Q = runoff
E = evaporation
G = change in ground water storage
S = change in seasonal snow storage
I = change in perennial snow and ice storage in glaciers

In many instances G will be negligible if a long sequence of years are chosen. Except in heavily glaciated catchments such as the Waitaki, I is unimportant for most New Zealand large rivers (Anderton 1973). Analyses using equation (1) are approximate and subject to some error, mostly in estimation of catchment precipitation (there are few gauges above 1000 m), and also in assessment of evaporation.

Snow and ice storage exert a significant regulatory effect on water yield from the Ivory Basin. Using equation (1), Anderton (1976a,b) estimates that snow melt contributed 21% of runoff during 1971-75, including a 9% contribution from melt of perennial snow and ice. This proportion is also similar in large catchments used for hydro-electricity generation such as the Waitaki. Anderton (1974), after examining mean monthly water balances for 1944–64, suggests the average quantity of snowpack stored each year in the Pukaki Basin is about 800×10^6 m^3 water equivalent, or 570 mm over the catchment area. This is sufficient to supply 20% of the mean annual lake inflow of 129 m^3 sec^{-1}. In an alternative analysis I. G. Jowett (pers. comm.) has shown storage to be 24% of runoff in the Lake Pukaki catchment for the period 1926-64, and 30% of runoff in the Hooker catchment for the period 1961-1969.

A similar water balance analysis has been completed for part of the Clutha catchment by Jowett and Thompson (1977), and is shown in Table 3. The accumulation begins in March, when the snow cover is at a minimum, and the snow melt season usually begins in October, although warmer spells of weather at any time during the winter cause some thaw. The average maximum accumulation of snow occurs in September, and is 225 mm or 14% of catchment runoff.

TABLE—3 Analysis of average annual seasonal variation of Clutha Lakes inflow 1933-75. After Jowett and Thompson (1977).

Month	*Catchment Runoff (mm)*	*Catchment Evaporation (mm)*	*Catchment Precipitation (mm)*	*To Storage (mm)*	*Accumulation (mm)*
					74
January	174	108	221	–61	13
February	156	77	220	–13	0
March	163	67	247	17	17
April	155	39	236	42	59
May	145	24	213	44	103
June	98	16	171	57	160
July	93	24	152	35	195
August	98	30	165	37	232
September	142	49	214	23	255
October	193	70	232	–31	224
November	210	88	237	–61	163
December	203	108	222	–89	74
Total	1830	700	2530	0	

Analysis of runoff records

As discussed, plots of daily flows of many South Island rivers with substantial parts of the catchment above 1000 m show an upward bulge in base flow in spring and early summer. This bulge can be interpreted as in Fig. 7.

Analysis of maximum yearly recorded Clutha lake inflows for the period 1931-75 shows that in more than half of the years the 4 day maximum inflow and 8 day maximum inflow occurred in the spring months. In 20% of the years they occurred in October.

Direct measurements

Toebes (1972) suggested the maximum volume of seasonal snow could be estimated by assuming half the South Island was covered with an equivalent water depth of 100 mm. This is probably too low, but gives a water volume of 7.7 km^3 or 3% of total runoff, estimated as 272.1 km^3.

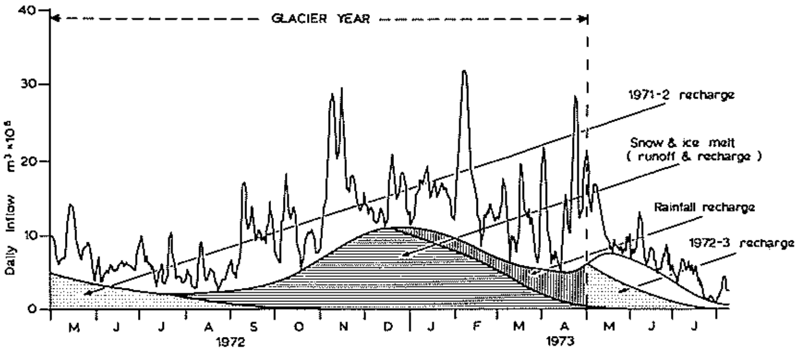

FIG.—7 Daily inflows to Lake Pukaki for the period May 1972 to July 1973, showing interpretation of base flow. After Anderton (1974).

Few regular measurements are taken of snow accumulation in New Zealand mountains. Data from Coronet Peak (Fitzharris, 1976) have been used to construct a simple model for the Clutha (Fitzharris, 1977). The seasonal snow line was assumed to be at 1000 m, and snow depths to increase, wedge-like, to 2.0 m at 2000 m, and thence remain constant to the permanent snow line at 2500 m. This is a very conservative estimate considering the very high precipitation close to the Main Divide (8000 mm).

This hypothetical snow wedge for the Clutha was then combined with hypsometric curves to compute the total water stored as snow in the Clutha catchment to be 3.5 x 10^9 m^3. This is equivalent to a river flow of 111 m^3 sec^{-1} for the whole year, or 23% of the mean flow at Roxburgh. However, most of the snow melts in the 4 months October to January, when the mean flow at Roxburgh is 573 m^3 sec^{-1}. The snowpack is equivalent to a flow of 445 m^3 sec^{-1} for these four months, or 77% of the Clutha flow at Roxburgh. Of course not all the water stored as snow reaches the river. Some is sublimated or evaporated, and some goes into recharging groundwater. Jowett and Thompson (1977) indicate that Clutha direct runoff is one half to two thirds of storm rainfall. Therefore, assuming half the snowpack eventually reaches the river, again conservative since the catchment is 'wet' during snow melt, then more than one third of the flow at Roxburgh in the spring and early summer comes from seasonal snow.

This review of the seasonal snow resource suggests that for South Island rivers with high catchments, snow storage is significant and can supply from 10% to 25% of annual flow. This conclusion is of interest particularly to the Clutha and Waitaki catchments, since both are important for electricity production.

Variability of Seasonal flow

When economic use is made of water from a hydrologic system for which little information on snow cover is available, it is not the average snow storage that is critical, but the variations about the average. Little is known of the inter-annual variability of snow cover in New Zealand, although it is thought to be large. From overseas measurements (Fitzharris, 1975) seasonal snow storage could vary by ± 50% from year to year. Because of the shape of the hypsometric curves, the size of the seasonal store in the Waitaki and Clutha catchments is more sensitive to changes of snow accumulation at elevations below 1500 m than at those above 2000 m. In the Ivory Basin annual melt estimated from the hydrograph varied between 18% and 24% of runoff over a 4 year period (Anderton, 1976b). Amount of melt varied between 1.97 m and 2.45 m. Jowett and Thompson (1977) show that the monthly mean flow of the Clutha river at Roxburgh tends to peak in November–December (median flow of 640 $m^3 sec^{-1}$), but between 1930 and 1975 varied from 370 $m^3 sec^{-1}$ to 1400 $m^3 sec^{-1}$, presumably reflecting in part the role of a variable snow cover. Weather also has a strong control on the timing of spring snow melt and of first freeze up, although no study of these effects has been made. At the end of individual snowmelt years the accumulation remaining in the Clutha lakes catchments varied $\pm$ 36 mm, or 14% of the mean annual accumulation (Table 3).

Comparison of Seasonal Snow with Hydro Storage

Anderton (1974) estimates the average quantity of the snowpack stored each year in the Pukaki Basin is equivalent to the power storage of Lake Pukaki (before it was raised in 1977). The snow store of 3.5×10^9 m^3 estimated for the Clutha is 1.5 times the size of the storage available in Lake Hawea (2.4×10^9 m^3, assuming a 20 m range of water level). Thus the seasonal snow store is comparable in size to the lake stores which the New Zealand Electricity Department monitor on a week to week basis.

While the seasonal snow store is unable to be controlled in the same way as a storage lake, it nevertheless is a large potential energy source which varies annually. The Clutha seasonal snow store from Table 2 is sufficient to generate 420 GWh of electricity annually in 1978, and about 770 GWh when DG3 and Luggate dams are completed in 1990. There is therefore a good case for monitoring the size of the snow cover, at least on a year to year basis, in the Waitaki and Clutha, and possibly the Waiau catchments.

MEASUREMENT OF SNOW

A wide variety of techniques are available, but some are unsuitable for the rugged mountains and high degree of variability induced by wind drifting that are typical of New Zealand conditions. There are two important aspects to be considered in assessing snow accumulation: one is the type of measurement technique to be used; the other is the problem of how to sample over space and time. This latter problem depends to some extent on the measurement technique.

Index Snow Courses

This measurement method has long been used in North America. Snow depths and cores are usually taken along sampling points on a line at the same location each year. Different types of snow sampler have been used in New Zealand in an attempt to find the one most suited to our conditions. In their early work on the Craigieburn snow courses Morris and O'Loughlin (1965) used an Italian tubular sampler. Chinn (1969) used a similar instrument in the Waitaki catchment. More recently the Federal or Leupold 'Mt Rose' sampler has become the standard instrument. Work *et al.*, (1965) estimate this sampler overestimates water equivalent by about 7% in light shallow snow, to about 10% or more in deep snow of high density.

Sampling difficulties are often caused by ice layers, high density snow or uneven ground conditions. Ice layers also create difficulties in interpreting summer surfaces on glaciers. To overcome this problem without digging pits, Chinn (1969) installed a cylindrical corrugated iron shaft on the Tasman Glacier. This enabled snow layers to be marked and sampled at any depth within the shaft.

Many factors influence the choice of location for snow courses. The site should be as representative as possible of the surrounding area, there should be a minimum of snow drift, the course should be readily accessible, and safe from avalanches. Often these requirements are difficult to obtain. At each snow course a number of different techniques have been used to locate sampling points. Chinn (1969) marked sampling points with permanent poles arranged in a straight line up the centre of the catchment. The Round Hill snow course had an additional traverse along the crest of the catchment. Archer (1970) and Morris and O'Loughlin (1965) took observations along the local contour. Gillies (1964) established a long, 19 km snow course in the Central Otago Fraser catchment; stations, which were cruciform about a central snow pole, were established at 0.8 km intervals. The poles were colour coded in black, red and orange bands to facilitate aerial inspection.

Such snow courses do not assess the water equivalent of the snowpack on a catchment basis, but attempt to provide an index of the snow cover, which can be related to runoff. After a number of years of observations, an empirical relationship between the snow course data and runoff can be established and used to predict the quantity of water from the spring melt. This concept has been most successful in North America (U.S. Army 1956) but is less likely to be effective in New Zealand. Fitzharris (1972) argues that the changing shape of the snow wedge, and resultant elevation–time interactions render unsuitable the snow course if measurements are taken at only one elevation on the mountain. More reliable indices will be obtained with a series of snow courses at different elevations, together with observations of the snowline.

The snow course concept requires a long period of record to establish an effective relationship with streamflow. This is not practical for New Zealand where many critical hydrological areas have no snow measurement records. Further, the snow courses provide a good index of catchment snow cover in continental North America because there is a clear accumulation and melt season. In New Zealand this is not so and therefore it is difficult to know when to sample in order to measure the maximum accumulation.

Sample Snow Courses

These differ from snow courses in North America in that several are often located on the same mountain, at different elevations, and on different aspects. They attempt to estimate the water stored as snow in a catchment, rather than index the snowpack. To some extent this approach, which is used in the U.S.S.R., and Japan, overcomes many of the difficulties of the previous method. Important questions are to know where, and how many, samples should be taken. Fitzharris (1976), using detailed snow depth and water equivalent observations on Coronet Peak, suggested that provided the snowline and the slope of the snow wedge above it are sampled at a 200 m elevation interval, then the shape of the spring snow wedge can be well defined with one further sampling site at higher elevation. In wetter regions, several large steps can develop on the snow wedge (Fitzharris, 1976) and more frequent sampling is required.

Little attention has been given to the number of snow depth or water equivalent measurements required to obtain good estimates of the population mean at any particular elevation. Again, using Coronet Peak data, mean snow depth can be estimated to $\pm$ 0.1 m at the 95% confidence level with from as few as 3 measurements at low elevation sites, where melt is well advanced, or at wind free sites, to as many as 100 measurements at exposed higher elevation sites. Somewhat similar sample sizes were obtained for the number of water equivalent measurements required, a daunting task if estimates are required for a range of elevations and aspects.

A compromise approach is to take the required number of depth measurements, and estimate the mean water equivalent by multiplying the mean of a number of snow density measurements obtained with the Federal snow sampler. Since snow density is a more conservative parameter over space and time than depth, the number of density samples required is usually less. Bartos and Rechard (1975) calculate the proportion of snow depth to density samples as the ratio of their respective coefficients of variation. Application to Coronet Peak data gives 1:2 to 1:7 for this ratio. Hence a scheme taking, say, 6 density samples and 30 depth measurements should adequately sample snow storage at the local scale. Obviously, such a sampling scheme is impractical for large catchments such as the Clutha or Waitaki, where there is great diversity of terrain. All that can reasonably be achieved here is to attempt to sample snow storage in a representative series of small catchments as an index of snow conditions in the larger catchment.

On the 0.8 km^2 Ivory Glacier, random and regular pole arrays were found to be equally efficient. A line of poles transverse to the glacier was of greater sampling significance than a longitudinal central line. Seven to 15 poles were adequate to measure snowpack ablation, with each pole estimating with an accuracy of $\pm$ 10% over an area up to 600 m^2. Accumulation was spatially more variable, with 30 poles probably necessary for estimation (Anderton and Chinn, 1978).

Size of sample can also be reduced with stratification into homogeneous subpopulations by examining where major variations in snowpack occur. On Central Otago mountains much variability is induced because of snow drifts behind vegetation, at breaks of slopes, in gullies, or in the lee of tors. These constitute natural traps for blowing snow which is driven across extensive undulating mountain areas by persistent south to west winds which may average 6–11 m/sec over a month (Bliss and Mark, 1974). Each type of trap can then be sampled in turn, in a stratified random sampling scheme. A drift at Duffers Saddle, on the Carrick Range, trapped 6 m^3 of water per linear metre of trap at right angles to the wind after one snow storm. By spring the volume of water had increased to 4000 m^3, or 27 m^3 of water per linear metre of trap (Fitzharris 1976). The areas between drifts had been blown nearly bare of snow. Therefore, on Central Otago mountains, any scheme to assess seasonal snow storage must consider measurement in potential drift areas.

Continuous Remote Measurement

The usual methods are to use a snow pillow, or radioactive gauge. These give the water equivalent of the snowpack at a site. Results can be recorded on charts or telemetered to a base station. These are not yet in use in New Zealand, although installation of snow pillows is being investigated at several sites. They have proved valuable overseas, but cannot overcome the spatial sampling problem.

Radioactive Profiling Gauges

These devices sample individual layers within the snow as well as giving the total water content of the snowpack. However none exist in New Zealand. They are expensive and do not overcome the spatial sampling problem, but are particularly valuable for predicting floods from rain or snow events when the snowpack is ripe, and for studying snow stratigraphy.

Aircraft and Satellite Observations

Aircraft observations have been used in Central Otago by Gillies (1964) and over the Tasman Glacier by Anderton (1975). They show promise for New Zealand conditions where the data base is sparse, and four or five flights a year can give a catchment-wide picture of snow cover behaviour. They cannot however give information on the water equivalent of the snowpack, or shape of the snow wedge.

Satellite observations have the same advantages and disadvantages of aircraft observations. Landsat photographs give high resolution photographs, where the snowline can probably be interpreted to $\pm$ 30 m elevation or better. However, the satellite makes a pass only once in 18 days, and even then the mountain areas may be covered by cloud. The DSIR Physics and Engineering Laboratory and the N.Z. Forest Service are investigating Landsat assessment of snow cover of a small basin near Lake Rotoiti. They are interested in differentiating various snow types from Landsat and aircraft imagery, and then assessing the area covered by each type. Research could be extended to assess a snow to water yield study (I. L. Thomas, pers. comm.).

Hickman (1972) has estimated the snowline for the South Island snow regions specified in Fig. 1 from meteorological satellite photographs for 1968–71. These have the advantage of daily observation, but have poorer resolution. Satellite camera resolution is of the order of 3 km at picture centre, and 8 km near its edge. In some cases snow line estimates can be made to within $\pm$ 75 m elevation, but generally are within $\pm$ 150 m elevation. A good 'reference' satellite picture of the snow covered mountains on a clear day aids in identification of ranges when there is partial cloud cover. An estimate of the snowline is made by comparing the relative width of a valley and adjacent mountain range with the same features on a series of contour charts. It is necessary to compare valleys on either side of a range with the range itself, since snowlines were observed to differ on slopes with different aspects.

In most countries a combination of the above methods has been found useful together with modelling procedures. No one method is sufficient alone. In New Zealand, there is currently no systematic measurement network to monitor the snow resource on a year to year basis. It is recommended that initial estimates should try to index the snowpack, rather than assess the actual total catchment water content, largely because of the difficult sampling problems. Any index should attempt to assess the elevation of the snowline, the area covered by snow, and the shape and behaviour of the 'average' snow wedge for the catchment.

Modelling the Snow Cover

Because there have been few measurements of snow in New Zealand catchments over a long period, there has been little opportunity to develop models that are a feature of North American snow hydrology. As well, there is often difficulty in obtaining reliable stream flow data and in separating the snow melt component of runoff.

Fitzharris (1972) has examined the possibility of establishing workable, empirical relationships between seasonal snow accumulation and elevation. Because the shape of the snow wedge is not always consistent, it is considered that simple relationships are not likely to produce reliable predictive models. A physically based approach might be more successful by including variables that are directly related to the processes operating. Proposed models should simulate, on a day to day basis, the variations in elevation of the rain/snow boundary, the amounts of snow deposited, and snow melt. Such an approach has been successfully applied to the Willamette Basin, U.S.A., using the SSARR snow melt watershed model by Anderson and Rockwood (1970).

Anderton (1974) and Jowett and Thompson (1977) have used the water balance approach to model the snow pack accumulation and depletion. Anderton interprets the storage term as indicative of snow storage in winter if the base flow is subtracted, and indicative of snow depletion in spring, when base flow is mainly produced by melt. An estimate of base flow can be obtained from the minimum daily flow for each month (Fig. 7).

Anderton also used indices of snowfall and melt to validate the water balance model for the Pukaki catchment. A snowfall index was computed by partitioning daily precipitation between snow and rain using a regional estimate of freezing level, and assuming that the rain/snow boundary is 300 m below the recorded freezing level. A melt index was computed as the product of the area of snow and ice below the freezing level, and the mean temperature for the elevation involved, using a lapse rate of 0.0075°C m^{-1}. A regional index for parts of the

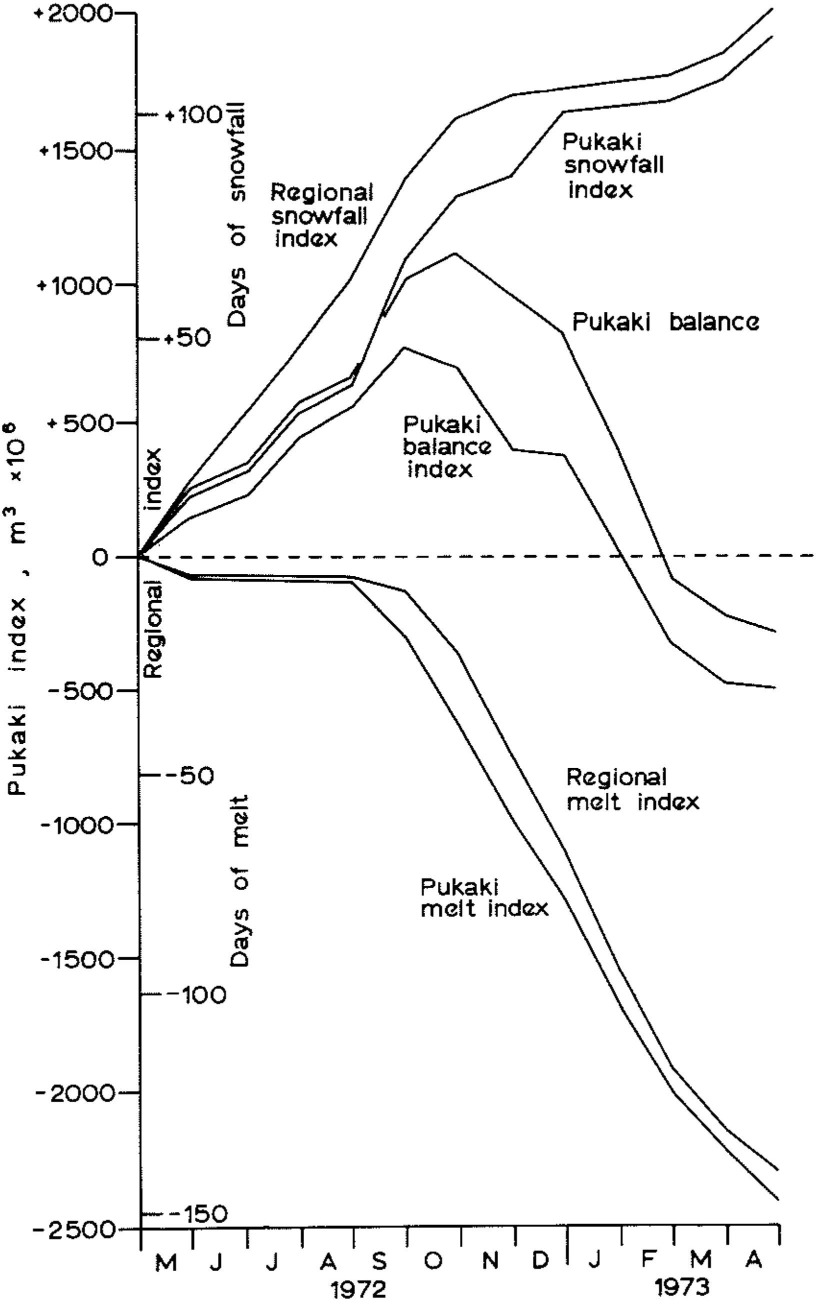

Fig.—8 Example and comparison of cumulative monthly snowfall, melt and balance indices for Pukaki basin, and cumulative monthly snowfall and melt indices for Southern Alps region (Mt Cook area) for 1972-73. After Anderton (1974).

Southern Alps was computed by assigning a value of +1 for each day when snowfall occurs, a value of −1 to each day favouring melt, and a value of 0 to each day when no significant melt occurs. The evaluation was based on daily weather maps and meteorological data. These simple indices show some promise, (Fig. 8) and should be pursued further for other years and other catchments.

When sustained measurements of snow accumulation are available in New Zealand there are good possibilities for using a wide variety of techniques to estimate snow melt. Kuzmin (1961) gives a detailed assessment and methodology for estimating the components of the energy balance over a snow cover, based on extensive Soviet research. A series of nomographs allow convenient calculation of daily melt rates. Empirical melt equations are widely used in the U.S.A., (Ven Te Chow, 1964; U.S. Army, 1956). However, most are for forested watersheds, but express daily melt as a function of temperature, an easily obtained parameter. More recent North American computer models also simulate, with sparse data, the energy fluxes over the snow surface.

Many studies of drifting snow from cold continental locations show that with persistent wind, snow will continue to drift until it sublimates; until it is moved to lower warmer elevations, where it melts; or until the velocity of the wind is reduced in the lee of natural traps in the landscape, where the snow falls out of the airstream and is deposited (Tabler, 1975). Thus in parts of New Zealand where drifting snow is common, such as the mountains of Central Otago, a method to estimate the maximum size of the snowpack is to assess the number of natural traps, and estimate their capacity. This has been attempted by Fitzharris (1977, Table 4) but is based on sparse data. When the traps are full, snow blows downwind, and may be lost from the catchment.

TABLE–4 Estimated maximum snow storage capacity of Central Otago mountain terrain. From Fitzharris (1977).

Nature of store	*Snow storage (mm water equivalent)*
Vegetation traps	500
Behind tors	2
In gullies and behind breaks of slope	95
In incipient cirques of block mountain fault scarp	84
TOTAL	680

CONCLUSION

The seasonal snow store is as large as hydro lake storage in some South Island rivers, and its melt, along with that from perennial snow and ice, makes a significant contribution to spring and early summer river flows. Hence snow hydrology is of some economic significance for energy generation and for irrigation schemes. However, the size and release of the snow store varies from year to year depending on weather. No comprehensive monitoring of the seasonal snow store is undertaken in New Zealand, although some attempt is made to index the annual mass balance of our most significant perennial ice resource, the Tasman Glacier. Many methods for estimating snow accumulation and melt are available. There is need to test some of these in New Zealand conditions on a routine basis.

Increased efficiency of water usage will be possible if snow is monitored, particularly for seasonal operation of hydro reservoirs, and for optimising other electricity generation. In the longer term, we might consider managing the snowpack. Augmentation by seeding of winter orographic clouds holds promise for increasing energy output in the Waitaki and Clutha hydro systems, if the initial results from the Western United States prove well founded. Howell (1977) suggests that for favoured localities an increase of 30% in size of snowpack could be regarded as a generous, but not unreasonable expectation from a mature technology intensively applied. The exposed, undulating surfaces of the windy Central Otago mountains may also be prime sites for reducing sublimation and increasing snow storage by using snow fences as discussed by Tabler (1971) and Martinelli (1973). Snow fences show promise for increasing total water yields, delaying the timing of snowmelt runoff, and concentrating snow for subsequent management.

An important issue which has received little attention to date, is the effect on water yield from snow melt as tussock lands are converted to improved pasture or some other form of landuse. Tussocks are effective traps for blowing snow and their replacement with lower vegetation will increase sublimation losses and the amount of blowing snow arriving at major natural terrain traps such as gullies. The overall effects on stream flow are difficult to determine exactly, but there could be a reduction in total water yield and a change in timing of the snow melt discharge.

The temperature profile and structure of the snowpack can also influence the hydrological response of an alpine catchment. Little is known of these aspects of the New Zealand snow cover, despite the early work of Heine (1962), but because of their maritime location and high wind regime, our alpine snow covers may have a structure different from those discussed in the overseas literature.

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SOIL WATER: A CATALOGUE, REVIEW, AND COMMENTARY OF NEW ZEALAND RESEARCH AND KNOWLEDGE OF WATER IN NEW ZEALAND SOILS

J. P. C. Watt

ABSTRACT

The New Zealand literature reporting soil water and related studies is assembled and reviewed. A commentary on the status of our knowledge of water in New Zealand soils is provided against a background of current international understanding of soil water processes. Future directions and fields of emphasis are suggested.

CONTENTS AND RANKING OF HEADINGS

INTRODUCTION

Knowledge of water in New Zealand soils has accumulated slowly over the last 30 years through practical experience and applied research in the fields of irrigation, land drainage, soil conservation and land use hydrology. Recently, ecological studies of plant communities have included valuable appraisal and measurements of soil water.

In addition, the continuing work on the characterisation of New Zealand soils has placed increasing emphasis on soil physical properties and especially the hydrologic properties of soil.

Technical appreciation and understanding of the physical processes involved in water entry to, storage within, and movement through soils is of fundamental importance to a country where soil is.the basis of the economy. For example, assessment of the versatility of a soil to different management strategies always has a soil water component; water conservation economies require that there be a scientific basis to irrigation if the water resource is to be used rationally and to its optimum; water yield, water quality and flood protection considerations require knowledge of how the soil and regolith act as primary modifiers of the rainfall/runoff process.

In a review of physical hydrology in New Zealand it is therefore appropriate that we assess the information currently available and if possible attempt to show where future work should be directed. In this review the literature included relates only to New Zealand soils and New Zealand research, and comes from local and overseas journals as well as proceedings of conferences and symposia. Some unpublished material such as university theses and internal reports have also been included. While the international scene greatly influences New Zealand research in soil water, the world literature in that field (e.g. Amerman *et al*, 1975) is far too formidable and dynamic for inclusion here.

An indication of the quickening interest in soil water problems and processes may be gauged from the increasing frequency with which interdisciplinary symposia and conferences now include at least a special session devoted to soil water. The first conference devoted wholly to soil moisture was convened by the Department of Scientific and Industrial Research and held at the Dominion Physical Laboratory in 1954. In 1956 the New Zealand Society of Soil Science held a symposium on soil moisture, and since its inception in 1961 the New Zealand Hydrological Society has on various occasions held discussion sessions and workshops on soil water at its annual conferences. In 1973 the D.S.I.R. convened the second 3 day 'Soil and Plant Water' symposium involving 40 participants, and in 1976 a third symposium attracted 55 participants.

SOIL PHYSICAL PROPERTIES AND HYDROLOGY

Physical properties of the soil have a large influence upon the time and space variability of soil water content and soil water potential energy. These physical properties (bulk density, porosity, pore size distribution, particle size distribution, and related parameters) form the basis of all soil water studies either directly or indirectly.

Little progress in measuring soil physical properties was made until comparatively recently. Birrell and Packard (1953) published the first paper to review the overall physical character of a soil parent material in their assessment of New Zealand loess. Other similar studies followed (Packard, 1954, 1957; McDonald, 1955, 1961).

With the development of a soil classification, attention switched from parent materials to soil groups. Physical properties of 54 'reference' soils (Gradwell and Birrell, 1968) indicate trends that exist between soil groups. However, because of the limited number of sites and soils sampled, the data must be used with caution. In addition, the analytical methods then used resulted in some inconsistencies. Ratings for soil physical properties are suggested by McDonald and Birrell (1968); these however are subject to review.

Specific relationships between soil physical properties and hydrology are first discussed by Birrell (1962) in a brief review of total porosity, macroporosity, available water capacity and clay content. Hydrological characteristics of soils are also discussed by Campbell (1962) with respect to empirical estimation of floods. Detention storage of rainfall in North Island soils is reviewed by Gradwell and Jackson (1970). The detention capacity of a soil (the volume of air–filled pores at field capacity) is compared with rainfall depth–duration–frequency information. The same theme of soil response to storm rainfall is continued by Jackson (1973) for the yellow–brown pumice soils of the Purukohukohu catchment and the yellow–brown earths of the Taita catchments; the former are concluded to be highly sensitive to changes in land use because the condition of the topsoil effectively controls the extent to which a large storage capacity can be satisfied.

Physical properties and soil hydrology are topics included in the New Zealand Society of Soil Science's publications on the Soil Groups of New Zealand. Current knowledge of physical properties, and their relation to hydrology, are concisely summarised for the pumice soils (Jackson, 1974a, 1974b; Noble, 1974), in Part 1: Yellow–brown Pumice Soils. This volume is also a good starting point for evaluating an extensive literature on land use hydrology and erosion on the volcanic plateau (Ward, 1956; Blong, 1966, 1970; Healy, 1967; Selby, 1971, 1972). Part 2 of the series covers the Yellow–brown Sands, but the only soil hydrology information relates to the Pukepuke black sand of the Manawatu (Mark–Brown and Scotter, 1977). Also included is moisture depletion data for *Pinus radiata* grown on the northern yellow–brown sands (Jackson, 1977). Part 3 of the series is in preparation and will cover gley soils. While the series is most certainly a useful summary of existing knowledge, the hydrologist will be disappointed at the lack of quantitative information relating soil physical properties to aspects of the hydrologic cycle; the data available relate only to very specific areas.

Physical properties of soils in the context of the pasture ecosystem are discussed by Cutler (1966). In the context of land use, physical properties of loess soils are covered in a series of papers relating to the South Island downlands (Watt, 1972). Land use effects on soil micromorphology are discussed by Barratt (1968).

Soil physical properties may be estimated as part of the field description of soil. The quantitative estimation of soil total porosity and macroporosity is considered feasible for New Zealand soils by McDonald and Julian (1965) who suggest methods and provide photographs as aids to such estimation. The estimation of permeability and infiltration from soil morphology is described by Griffiths (1975); work continues in improving the criteria and their interpretation. Water–table regime and permeability are integrated in the assessment of drainage classes of soils (Taylor and Pohlen, 1962).

MEASURING SOIL WATER

Moisture Content and Moisture Tension

Useful general reviews of the international literature, accompanied by remarks on local New Zealand conditions and experiences, are given by Noble (1956, 1973) and Blake (1973).

The thermo–gravimetric method is covered comprehensively by Reynolds (1970) who takes into account the many factors with which field workers are familiar. Sampling errors associated with the method are also discussed by Fitzgerald *et al.* (1963). The water content obtained is on weight basis and may be expressed as the dimensionless ratio of weight of water lost on drying divided by the weight of oven–dried soil (moisture weight fraction, MWF). If the water content is required on a volume basis (moisture volume fraction, MVF), the volume from which the sample was taken must be known, which introduces a further source of error.

The neutron probe is now regarded as standard field equipment for measuring water content on a volume basis. In 1977 over 30 instruments were registered with the Department of Health's National Radiation Laboratory. Noble (1958) and McLellan (1974) review the principles involved and the development of the method.

Five years after United States and Canadian scientists first reported the technique, D.S.I.R.'s Dominion Physical Laboratory had, in 1956, the equipment necessary to construct a soil–moisture meter using a radium–beryllium source and an enriched boron trifluoride counter as detector (McCallum, 1956). This was apparently tested at Makara (Wellington) as a surface detector and later as a probe in boreholes. Experience at Winchmore (Canterbury) using a commercial model based on the Dominion Physical Laboratory's design, but using cobalt 60, was reported by Fitzgerald (1963). Emphasis at this stage was on calibration as a surface probe rather than as a depth probe. Blake (1972) reports the first New Zealand assessment of the second generation probes in his study of sampling requirements of the Troxler instrument.

Calibration studies are reported by McLellan (1974) and Beets (1977), both using the Wallingford probe. Problems associated with calibration appear to include probe resolution, soil variability, and the amount of hydrogen present in the soil but not necessarily released during oven drying. A preoccupation with calibration by a number of workers whose work has yet to appear, has tended to obscure the great attribute of the neutron probe which is its ability to monitor temporal *changes* in water content throughout a profile using reasonable sample volumes. Methods of correcting readings at the surface are summarised by Painter (1977). Determination of the sensitive centre for use as the reference is described by Maber (1971) in the context of a study of positioning wetted depth fronts in trickle irrigation.

Resistance blocks, which measure moisture tension in the range 0.5 to 15 bars have been used in a number of studies and appear to have particular value in moisture regime studies (Jackson, 1966; Watt, 1977c). Rickard (1956a) found them unsuited to irrigation research where frequent wetting and drying hastened disintegration. Warren (1975) however has found them of general practical value as aids to scheduling irrigation. Closs and Jones (1955) reported early attempts at the construction and installation of gypsum 'soil moisture meters', and employed them to measure moisture losses in one of the early field studies of transpiration (Closs, 1956b). Recent reviews of the literature, and assessments of block construction, calibration and use are given by Prestidge (1972) and Ryu (1976). In comparison with gypsum units, Will (1959) found that fibreglass units gave a more rapid response to a lowering of the moisture content, and were also more sensitive to moisture changes at low tension. The use of fibreglass cells is also reported by Prestidge (1972) and Archer (1976).

The measurement of positive water pressures with piezometers is reported by O'Loughlin (1974), and small negative 'tensions' (0 to 0.8 bars) using tensiometers by Cook and Watt (in prep.). A capacitance probe especially suited for *in situ* moisture measurement of highway materials is described by Nguyen (1975), and its potential briefly outlined by Painter (1977).

Sampling Soil Water

Recent studies in effluent disposal by irrigation have required monitoring and sampling for soil water quality. In a lysimeter study of unsaturated leachate, Bell (1974) evaluates 1 bar porous–ceramic soil moisture samplers that are commercially available overseas. He found that the ceramic cup filtered out bacteria and were therefore only suited to studies of chemical changes. Quin and Forsythe (1976) describe 'all–plastic suction lysimeters' which are claimed to be suitable for all microbiological and chemical analyses without risk of contamination, and which can sample percolating waters to 9 m depth. Stevenson (1978) describes a simple and inexpensive suction sampler in which air entry and vacuum loss occur only at high soil moisture tensions and thus permit the sampling of soil water over relatively long periods.

SOIL WATER PROCESSES

Entry of Water to Soils

Methods of measuring infiltration or water entry to soil are of three types: hydrograph analysis, cylinder or plot ponding, and rainfall simulation. The watershed hydrograph method integrates the total infiltration for a catchment during a storm through comparison of rainfall and runoff (Toebes, 1962a). This single infiltration curve does not allow for the variations of slope, soil and vegetation that occur within the catchment. Toebes (1962d) found that Horton's infiltration equation gave the best fit to such field curves especially when the time factor was large; Kostiakov's and Philip's equations were valid for only limited time periods.

The ponding method involves impounding water within a plot or cylinder and measuring the rate of water intake. An example is the buffered double–ring infiltrometer which has been used in a study of infiltration properties of a steepland yellow–brown earth in Canterbury (Gillingham, 1964), and in irrigation investigations in Otago (Rickard and Cossens, 1966). Its use as a tool in the assessment of land for irrigation is advocated by Griffiths (1975). Both Gillingham (1964) and Fitzgerald *et al.* (1971) conclude the method unreliable. Problems associated with the method include disturbance to the soil on installation, air entrapment, seepage at the cylinder perimeter, lateral movement of water, and the small volume of soil actually sampled. The method's main value appears to be in distinguishing between broad classes of infiltration rather than in determining absolute values or minor differences.

The rainfall simulation method simulates natural conditions on a small plot, the difference between the rates of application and runoff being taken as the infiltration. Studies using the North Fork infiltrometer are reported by Nordbye and Campbell (1951), Campbell (1956) and Gillingham (1964). Blake *et al.* (1968) working at Puketurua in Northland, extensively modified the original design, but conclude in their evaluation of the method that it is costly and not recommended for widespread use. Selby (1970) modified another design for a portable rainfall simulator which did not disturb the soil and which had low water requirements, and found it "convenient, versatile, simple and relatively cheap".

The soil and plant factors affecting infiltration of soil have been discussed and evaluated in a number of these studies. They range from generalisations based on meagre data to quantitative assessments based on statistical studies. Most report high variability which is a reflection of the spatial variability of the soil–plant system, and the measurement method used. All studies agree with Campbell's (1956) conclusion that "in general, management exerts an overwhelming influence on infiltration rates and tends to mask vegetation and soil type differences".

Soil physical properties are undoubtedly important in controlling infiltration, but exact relationships are complex and difficult to define. On the yellow–brown pumic soils Selby and Hosking (1971) found that while statistical analyses reveal that infiltration correlates with several soil properties, the best prediction equation explained only 57% of the original variation. They found that infiltration was least into soils beneath pasture which had been compacted by land–use practice, especially when the soils had low pre–existing moisture. Erodibility of these soils is discussed by Selby and Hosking (1973), and runoff by Selby (1973). Also working on the yellow–brown pumice soils, Pittams (1973) provides the only New Zealand study of water movement during infiltration and redistribution, and this constitutes a useful contribution to our knowledge of moisture recharge in these soils. Lynch (1975), in a study of the effect of land–use and soil physical properties on infiltration in a yellow–brown pumice soil, concludes that the complex inter–relationships between soil properties and with land–use, together with seasonal variation in moisture content, makes prediction of infiltration rates almost impossible. Working on a yellow–grey earth in Otago Dons (1976) emphasises the variability of infiltration capacity as a function of real spatial variability of the soil system. Fitzgerald *et al.* (1971) in a study of 21 Otago soils derived from greywacke and schist report very little association between parameters of fitted infiltration equations and the soil physical properties expected to exert an effect on the process.

Water repellancy, or resistance of soils to wetting, has not attracted much work in New Zealand despite the fact that the problem is present, has important land–use connotation, and is a field of active research overseas. In the yellow–brown pumice soils near Taupo, Van't Woudt (1959) concludes that under certain conditions an improvement in the base status of the soil is associated with improved wettability, and that in some cases particle coatings by hydrophobic films repel water. Will (1962) notes that intense fungal activity around roots of *Pinus radiata* result in pockets of bleached soil which are resistant to wetting and which dry to higher moisture tensions. The pockets are associated with complete canopy cover and a thick litter layer and disappear when the forest is cleared. John (1978) has recently demonstrated that heat can induce water repellancy in pumice soils and therefore fire may be a significant factor.

Forest floor hydrology is a long neglected area of study in New Zealand and relates directly to the entry of water to soils. In forest, the distribution of rainfall at the soil surface is affected by the varying proportions of stemflow and throughfall. Aldridge and Jackson (1973) suggest that stemflow water has an important influence on soil properties and moisture distribution, but there are no studies which assess such spatial variability. In the first serious study of the physical nature of the forest floor, Webster (1976) describes conditions and variability under beech–podocarp forest in north Westland and Nelson. Some preliminary physical and hydrologic properties of the litter and humus layers are presented.

Movement of Water within Soils

Theoretical studies

Theoretical aspects of soil water movement that have been studied under New Zealand auspices are outlined in the following references. These should not be divorced from the international literature. Segedin and Miller (1962) have published a two dimensional analysis of the 'standpipe' method of determining soil permeability, by solving an analogous problem which considers flow from a lined ditch into an infinite homogeneous soil. Miller (1962) extends these results to take account of a horizontal sand layer in the flow.

Further into the field of fluid mechanics, Henderson and Wooding (1964) study the 'buildup and decay of a laminar or turbulent flow over a sloping plane' and extend the problem 'to include the case of groundwater flow through a porous medium overlying a sloping impermeable stratum where water is supplied by infiltration from the groundsurface above'. McNabb (1962) applies Wooding's equation to a horizontal situation, noting it to have some bearing on problems concerned with drainage, irrigation, and diffusion of tides into porous rock. Unsaturated seepage flow from a horizontal boundary is discussed by Wooding (1975) in a mathematical study of relevance to both infiltration studies and to the analysis of the movement and dispersion of nutrients or pollutants. A review of theoretical soil water problems including capillary uptake, the physical effects of pore geometry, and the macroscopic flow equations is given by Wodding and Morel–Seytoux (1976). A theoretical infiltration model for sandy soils is summarised by Wooding (1973).

More recently Clothier *et al.* (1977a) use simplified theory to develop a model relating the drainage flux at the base of a soil column underlain by a coarse–textured layer to the water storage in the soil. The decline in the drainage flux with time after wetting is shown to depend on the depth and retentivity of the soil and the hydraulic conductivity of the underlying coarse layer. Use of the model requires only simple field measurements. A simplified model of water movement in the soil profile has also been used by Stevenson and Wilcock (1977) for calculating concentrations of chemicals in samplers following effluent application.

Measurement

Measuring the rates of movement of water in soil has likewise received scant attention, although some recent work suggests increasing interest in this important field. Using laboratory methods, Gradwell (1974a) measured unsaturated hydraulic conductivities at water contents near field capacity, for a range of subsoil horizons. He compared the measured values with calculated conductivities using Marshall's (1958) method and found considerable differences. Although a considerable range of matching factors were found he considers 0.07 times the calculated hydraulic conductivity value gives a sufficiently accurate value for comparing soils as suppliers of water to roots. Calculated hydraulic conductivities for granular brown loams and northern yellow–brown earths based on pore–size distribution data are discussed further by Gradwell (in press).

Saturated hydraulic conductivity has been measured in the laboratory by Harvey (1974) and Corker (1977) using constant head permeameters and undisturbed soil cores. Harvey found trends which could be expected in terms of other measured properties and derives a relationship with bulk density to predict saturated hydraulic conductivity. Corker found the method difficult to use for the dense B and C horizons of a yellow–grey earth and results were often unreliable due to air entrapment, clogging of pores through biologic activity, and the dominating influence of cracks and holes on the flow process.

Problems associated with taking small samples from the field to the laboratory where they are separated from their continuum and inevitably suffer from some measure of disturbance, are overcome by field methods for measuring hydraulic conductivity. Improvements in field techniques have been reported in the overseas literature over the last decade and these are now being looked at in the New Zealand situation.

Measurement of saturated hydraulic conductivity (Ksat) above a water table using the 'shallow–well pump–in' method is reported by Noble (1977).

Saturated hydraulic conductivities below a water table are reported by Corker (1977), and Mark–Brown and Scotter (1977). The 'auger hole' method used by Corker was found to be quick, easy and reliable when limited to topsoils, and was assessed as best suited to homogeneous non–layered soils where Ksat in the horizontal direction is required. The 'piezometer' and 'tube' methods also tested by Corker differ only in the shape of the cavity at the base of the small–diameter pipe. This essentially determines that flow should be in a vertical or horizontal direction respectively. As noted by Corker (1977) the main advantage of these latter two methods over the 'auger hole' method is the improved resolution of the measurement depth.

A 'transient drainage' method for the field determination of unsaturated hydraulic conductivity has recently been tested and evaluated (Cook and Watt, in prep; Cook *et al.* in press). The method requires coincident determinations of water content and matric tension in a draining profile where water movmement is assumed in the vertical direction.

The percolation of water and contaminants through soils to groundwater is another topic which must receive more attention in the future, especially in the context of the impact of irrigation schemes and effluent disposal systems on the environment. A recent study (Thorpe, 1977) describes efforts to evaluate the potential hazards of urbanisation over an unconfined aquifer at Hastings. Penetration of pollutants into the aquifer, groundwater velocities, and dispersion characteristics are investigated. Salinity and sodium problems can arise with irrigation if groundwater levels are high and if the groundwater is saline. Groundwater and soil salinity are examined for the Kurow–Duntroon district in the Waitaki Valley (Cossens and Rickard 1970b) and in the Ida Valley, Central Otago (Cossens and Rickard 1968). In both cases it is concluded that areas of salty soils could be leached satisfactorily by controlled irrigation.

Storage of Water within Soils

The soil moisture characteristic is the relationship between matric suction and water content. This curve, variously known as the pF curve, water retention curve, or moisture release curve, is usually determined in the laboratory on 'undisturbed' cores. It replaces the older concepts of 'hygroscopic', 'capillary' and 'gravitational' water by representing as a continuum the relationship between water stored in a soil and the energy to be expended in extracting that water. Since all soil water phenomena are concerned with water content and energy gradients, this curve is of fundamental importance to studies of soil water statics and dynamics.

Soil moisture characteristic data have accumulated slowly from the late 1950's when Packard (1957) published data for the 'Taupo pumice sands' (albeit using dried seived samples from the soil surface). A systematic appraisal of the soil–moisture characteristic and water–holding properties of representative New Zealand soils used for pastoral farming was commenced by Gradwell (1968a) in a study of Waikato soils. This paper also covers methods of determination. The series continues with North Auckland soils (Gradwell, 1971), southern and central zonal soils (Gradwell, 1974b), and additional intrazonal soils (Gradwell, 1976), the last oeing a country–wide amplification of the earlier Waikato work. There are considerable differences between soils in the amount of water released between tensions of 0.2 and 15 bars. Subsoils show proportionately larger differences than topsoils. In most cases differences between soils can be explained in terms of the main soil groups. An exception are the central yellow–brown earths where elevated higher–rainfall members release much more water than other members. Ranking soil groups in order of decreasing subsoil–available water (0.2–15 bars) indicates: upland central yellow–brown earths > southern yellow–brown earths > yellow–grey earths, northern yellow–brown earths, central yellow–brown earths > brown loams, brown granular loams. A three fold difference separates the first group from the last (Gradwell, 1973).

Further studies of differences between soil groups, in terms of pore size distributions in the 0 to 1 bar range of drainage tensions for different horizons, indicate the yellow–brown loams to be generally more porous than other soil groups studied; differences between other soil groups are mainly for particular pore sizes at particular depths (Gradwell, in press). It was also found that in most samples, pores with diameters greater than 60 μm (0–0.05 bar tension equivalent occupy a greater volume than pores between 60 and 3 μm 0.05–1 bar). Most soil groups have similar pore size distributions in the 60–3 μm range, but a few groups such as the yellow–brown loams, their intergrades with yellow–brown earths, and the central upland yellow–brown earths have 'exceptionally' large volumes of pores close to 60 μm in diameter.

The moisture characteristic for a range of subalpine and alpine soils is given by Archer (1976). A considerable range of storage values between steepland, gleyed, and organic members are noted. Data from Molloy's (1964) and Macdonald's (1961) site were considered similar to Archer's steepland member.

If a distinction is to be made between the moisture characteristic of a soil and its available water capacity it is that the latter is an agronomic appreciation of the potential of the soil to store water in the plant rooting zone whereas the former is a physical definition of the water content/tension relationship for all horizons for all moisture contents. Much of Gradwell's work is used to estimate available water capacities, and in the only other major study the emphasis is the same although the orientation is somewhat different. This other study is the series of papers covering irrigation investigations in Otago where available water capacities and other physical properties are determined for the major soil series of the Ida Valley (Rickard and Cossens, 1966), Arrow Basin and Upper Clutha Valley (Rickard and Cossens, 1968), Maniototo Plains (Cossens and Rickard, 1969), Kurow district (Cossens and Rickard, 1970a), Lower Waitaki Plains (Cossens *et al.* 1971) and Manuherikia Valley (Rickard and Cossens, 1973a, 1973b). The New Zealand Standard 5103 on sprinkler irrigation gives available water capacities of soils on both a texture basis and a soil group basis (Standards Association of New Zealand, 1973). The data is derived from the above mentioned references. Future irrigation manuals will no doubt expand this information which should only be used in the absence of specific site data.

Available water capacity is a soil property often referred to and occasionally measured. Its definition however is not always clear. It refers to the volume of water stored in the soil and available to crops, and so is an integral of moisture over the rooting depth. It also refers to the range of water contents above which the presence of water is transient due to drainage, and below which plants are unable to extract water.

The definition of rooting depth presents problems. Plant species differ in their root distribution, the depth of soil explored, and their seasonal water withdrawal patterns. Soil survey practice is to take it as synonomous with depth to impermeable layer (Griffiths, 1975), or as that depth to which roots are observed. Aspects of rooting depth are discussed for lucerne (Cox, 1967), for some crop and pasture species (Evans, 1977, 1978), and for *Pinus radiata* in pumice soils (Will, 1965; Will and Stone, 1964, 1967). A general discussion of plant root distribution and soil water extraction is given by Evans (1973). There is little doubt that this field will attract considerably more interest in the foreseeable future, as soil physics and agronomy become more closely related. Root survey methods are reviewed for hydrological experiments by Blake and Branson (1969).

The lower limit of available water is generally accepted as the 15 bar value which corresponds approximately with the permanent wilting point of most crop plants. Although there are good reasons why 'wilting point' should not be regarded as a soil constant, it has many practical uses.

The water content taken as the upper limit is quite variable. Laboratory workers frequently use either 1/3 bar or 200 cm water tension values. Field workers use 'field capacity' which may be either the water content after 48 hours of natural drainage following saturation, or the moisture content prevailing when the drainage curve flattens out. The two are frequently similar, but in some freely draining soils appreciable volumes of water can be lost to drainage well after 48 hours. Lack of appreciation of this drainage factor can cause overestimation of available water. Artificial drainage in a paddock can alter the field capacity. Underlying horizons of different texture can also influence the drainage flux and therefore water retention. This phenomenon has been known for some time but has been given recent quantitative emphasis by Clothier and Pollok (1976), Clothier (1977a), and Clothier *et al.* (1977b). A timely and comprehensive discussion of field capacity and available water is given by Scotter (1977).

In hydrological studies it is sometimes convenient to consider water held in the soil above field capacity as detention storage (available to streams) and that held in at the lower moisture contents up to tensions of 15 bars as retention storage (available to plants). Studies which have taken particular note of the detention characteristic of soils in hydrologic assessment include Gradwell and Jackson (1970), Watt (1971), and R. J. Jackson (1973).

The retention of water in a soil in a pot will differ from the retention of that soil in the field. Concepts of water storage and retention in containers has been briefly mentioned by Scotter (1977), and is elaborated upon by Goh and Haynes (1977) and Haynes and Goh (1978) in an evaluation of potting media.

Water storage relationships in the greywacke screes of the South Island high country is touched on by Fisher (1952) and developed by O'Loughlin (1965) in his tentative classification of scree types based on geomorphic features and nature of origin.

Water Runoff and Erosion Processes

Concern for soil erosion in New Zealand (McCaskill, 1973) has led to considerable interest in water runoff processes, but few quantitative studies have been completed. Most studies (e.g. Selby, 1973) have been of surface runoff and its relation to sediment transport. Other forms of runoff involving sub–surface movement such as described by Van't Woudt (1954) have received less attention. This is unlikely to remain the case with the growing interest in 'variable source–area' concepts of storm flow (Hayward, 1977), and with evaluation of the relative significance of runoff types as sources of phosphate and nitrate enrichment to streams (McColl, in press; Rennes *et al.* 1977).

The runoff plot method for measuring soil loss has been reviewed and critically discussed by Hayward (1967, 1968) and Boughton (1967). The latter notes two problems that arise with the method: bias of the data caused by plot construction; and statistical analysis of the uncertainty associated with the results. The method has been used in New Zealand by Campbell (1945) in an early Hawke's Bay study of land deterioration; by Soons and Rainer (1968) and Soons (1970) to study erosion processes and rates on varying slopes and vegetation covers at Cass; by Selby (1973) in an investigation of causes of runoff from a catchment of pumice lithology near Lake Taupo; and by Duncan (1972) in an investigation of plot hydrology. Hayward (1969), in an attempt to determine the extent of soil movement within a small catchment in the Rakaia River headwaters, concludes the method as 'totally unsuitable' for erosion research.

In soils having relatively high saturated hydraulic conductivities and free drainage, rainfall intensity may rarely exceed the infiltration capacity and Hortonian surface runoff rarely occurs (Hayward, 1977). Soils having poor drainage on the other hand (e.g. the yellow–grey earths) may be near saturation for significant periods and at these times possess a storage opportunity too small to accommodate precipitation. In this situation surface runoff occurs in a process similar to that when surface compaction restricts rate of water entry, and surface–wash and rill–wash erosion may occur, especially on cultivated slopes (Raeside and Baumgart, 1947; Watt, 1972). Mole drainage has been shown to reduce surface runoff by 25–30%, especially during autumn and early winter; the practice is less effective in reducing surface runoff during prolonged rain (Rennes *et al.* 1977).

Tunnel–gully erosion (soil piping) is initiated by sub–surface dispersion and subsequent downslope transport of material. It has been described by Gibbs (1945) at Wither Hills, Marlborough, and by Hughes (1972) on Banks Peninsula. It is a feature of the drier yellow–grey earth environment throughout the South

Island. Laffan and Cutler (1977) give an up to date discussion on the mechanisms involved, noting dispersibility to be correlated with high exchangeable sodium and low organic carbon.

Mass movement may be also regarded as a soil–water phenomenon. An investigation into the mechanisms of soil slips in relation to soil properties and climate is reported by Jackson (1966). The assessment of pore water pressures as a factor in determining the stress on a potential failure plane is described by O'Loughlin (1974) in evaluating the effect of timber removal on the stability of forest soils. Regolith saturation, promoted by an underlying impermeable sandstone, is identified (O'Loughlin and Pearce, 1976) as a principal cause of instability of forest soils in north Westland except where additional shear strength is provided by tree roots. Earth–flows and their related environmental factors, including water content (surface only), depth, and sensitivity of the regolith, are described for eastern Otago by Crozier (1968, 1969). Mass movement is described and assessed in the now–closed Tangoio Conservation Reserve in Hawke's Bay by Eyles (1971). Painter (1973) describes the physical principles involved in mass–movement. Rates of soil creep are discussed by Selby (1974) for pumice soils and deposits. Greenland and Owens (1967) also describe soil moisture conditions and soil creep at Cass but conclude its overall contribution to be small compared to freeze and thaw action.

Evapotranspiration

This important process links the soil–plant–atmosphere systems. As such it extends well beyond the scope of a review concerned only with soil water, and the New Zealand experience is reviewed in a separate chapter of this volume. Link papers include those of Kerr (1973) who discusses evapotranspiration of pastures and crops, D. S. Jackson (1973) who discusses methods of estimating evapotranspiration by forests, and Kerr (1974) who examines some of the factors responsible for loss of water from soil by evaporation.

Frost Action Features and Processes

The subalpine and alpine zones of the New Zealand mountains have many periglacial landforms including solifluction lobes and terraces, soil stripes, polygons, and soil hummocks. This microtopography is considered a product of frost action, freeze–thaw activity, meltwater movement, and wind transport. Some features are products of past environments and therefore relic; others are actively forming. Their hydrologic significance, if any, has not been assessed although the snow detention properties of soil stripes and hummocks have been noted. McCraw (1967) has observed that some forms of patterned ground such as stone rings and stripes are more numerous and more clearly developed at the higher elevations in New Zealand than they are in Antarctica. Patterned ground and surface features have been described at Molesworth (Gradwell, 1957), in western Otago (McCraw, 1959), on the Old Man Range, Central Otago (Billings and Mark, 1961), and in Antarctica (McCraw, 1967). A general survey of periglacial features in New Zealand is given by Soons (1962). Microtopography is also referred to in relation to the high alpine vegetation of Central Otago by Mark and Bliss (1970). Preliminary investigations into a form of miniature stone stripe in east Otago are reported by Brockie (1967). These references do not provide an exhaustive review but serve as an introduction.

There are three categories of frost–action processes: frost shattering of outcrop rock (gelifraction); flow or creep of seasonally saturated material (gelifluction); *in situ* redistribution or reorientation of particles by frost heave (cryoturbation). In the South Island high country, cryoturbation has been of particular interest because of its influence on grassland establishment and maintenance following burning of the original tussock grassland cover and its associated litter. Concern for erosion in the high country led to Gradwell's early studies on fescue tussock grassland at Molesworth (Gradwell, 1954, 1955) and on snow tussock country at Fox Peak in South Canterbury (Gradwell, 1960b). His earlier study records forms and distribution of topsoil ice on a number of sites under varying weather and snow conditions. Four distinct ice forms were observed: vertical needle ice, flat horizontal lenses, small granular crystals, and continuous interstitial ice. Needle ice formation occurred most frequently in bare areas and appeared favoured by partial thaw during the day, high waterholding capacity, and inferior drainage. The powerful influence of needle ice in preventing seedling establishment is noted, as is its potential to undermine mature tussocks. The snow tussock study emphasises the insulating influences of tussock litter and scree in reducing frost action. Erosion by needle ice is further reviewed by Soons (1968) who reports as a consequence an inverse relationship between precipitation and sediment yield in runoff plots during winter. Observations on the growth of needle ice are reported by Soons and Greenland (1970).

The influence of frost on physical properties of the surface soil is pronounced and well known, but rarely quantified. In a comprehensive soil appraisal of a subalpine area, Harvey (1974) recognises several erosion variants of the Puketeraki soil association on the Dog Range (Canterbury), and concludes frost action to be the most important phenomenon affecting soil properties. He notes that through the comminution of rock particles, the destruction of soil structure, and the disruption of surface horizons, frost action affects hydraulic conductivity, bulk density, total porosity, particle size distribution and water–holding capacity.

There are no New Zealand studies on the hydrologic significance of frozen ground, despite its importance in inland areas of the South Island in winter. Cryopedologic processes have not been studied either.

THE WATER BALANCE

The field water balance is an accounting of the water status of the soil system, and its inputs and outputs. It represents the net result of the interdependent processes of water in the soil–plant–atmosphere system. On a catchment basis, water balances are of particular hydrological interest (Toebes, 1962b). On a site basis they can allow the analysis of the constituent processes (Clothier, 1977b). Most studies however pertain to climatological aspects rather than soil physical aspects.

Rickard and Fitzgerald (1970) review the considerable New Zealand literature on several aspects of the water balance. Their concern is with the Penman and Thornthwaite methods of estimating evapotranspiration and the relation of the soil water balance to irrigation, forecasting soil moisture, and plant growth. The following account updates Rickard and Fitzgerald's review.

Estimation of Evapotranspiration from Climatological Data

Coulter (1973a) considers that the method most applicable for broadscale estimates of evapotranspiration in New Zealand is to calculate potential evapotranspiration (PE) and then to assess the extent actual evapotranspiration (AE) falls below it as soil moisture becomes limiting. Methods of estimating PE, and their application to New Zealand conditions, have been discussed in surprisingly few papers. The Winchmore (Canterbury) experience indicates the Thornthwaite estimate (Toebes, 1968; Fitzgerald, 1969) to be equally satisfactory to the original Penman method based on open–water evaporation, and superior to the subsequent Penman modifications. This was for estimating soil moisture deficits, using gravimetric AE determinations as the basis for comparison (Fitzgerald, 1974). Calculated PE values were converted to AE using using an AE/PE relationship described by Rickard and Fitzgerald (1969). At Lincoln College, Heine (1976) found that Thornthwaite's estimate of annual total PE underestimated that of Penman's combination method. Coulter (1973a) found PE estimates by Penman's and McIlroy's formulae to agree well with 5 day and monthly 'open water' evaporation values calculated from evaporation tank data by Finkelstein (1973). Tank estimates however were greater for stations where the aerodynamic term was large. Thornthwaite estimates were considerably lower except where evapotranspiration was near that corresponding to wet conditions over a wide area. He concludes that on the basis of overseas testing over many years, Penman PE estimates should be reliable; Thornthwaite estimates probably underestimate the true summer values, seriously so in the drier eastern areas. Penman's method was first discussed for the New Zealand situation by Closs (1956a), and its ability to be adapted to the nature o site has been emphasised by Jackson (1967, 1973).

The shape of the AE/PE curve during soil drying is discussed by Toebes (1962c), Rickard and Fitzgerald (1969) and Coulter (1973a). An attempt to establish this from field data is described by Watt (1977b). Formulae are also available for estimating AE but they require detailed input and there has been as yet no attempt to apply them on a macro–scale in New Zealand. Coulter (1973a) therefore uses a water balance approach to determine AE. He assumes that when rainfall is much less than PE over a period, AE becomes dependent on the amount of available water in the soil, plus precipitation. In dry periods when the soil moisture store is depleted, AE depends essentially on rainfall alone and may be independent of PE. In computing broadscale indices of 'drought incidence' and 'irrigation water need' there is little need for an accurate knowledge of the AE/PE curve since plant production generally declines as soon as water deficits are sufficient to prevent PE being realised.

Soil Moisture Deficits and Water Need

Irrigation research in New Zealand has focussed particularly on the size and duration of water deficit and the consequent water need of various soils and crops. Most attention has been directed at the deficit under pasture, but some work is being initiated on other crops (see Stoker, 1977 and references therein).

Both the Winchmore Irrigation Research Station and the New Zealand Meteorological Service have reported substantial work on the determination of deficits. Early work on predicting soil moisture deficits at Winchmore (mid Canterbury) and Central Otago is reported by Rickard (1956b, 1957, 1960, 1961), Fitzgerald and Rickard (1960), and Fitzgerald and Cossens (1966). Rickard's study (1960, 1961) is of special interest as it was the first on drought at a particular station (Ashburton) where drought is defined with specific reference to the moisture state of the soil. Early discussion of irrigation and water need throughout New Zealand was published by Gabites (1956). Rainfall variability in New Zealand had at this stage been assessed by Seelye (1946), dry periods and periods of low rainfall by Kidson (1931) and the occurrence of drought by Bondy (1950). Maps of average daily evaporation losses for different months were first published by Gabites (1960).

Coulter (1963) assesses rainfall and water need, and gives water balance calculations for 13 North Island stations using Thornthwaite PE. Maps of New Zealand showing water deficit information using Thornthwaite PE were subsequently published (Coulter, 1966, 1973b, 1973c). Monthly water balance summaries to 1970 for several New Zealand stations, using Thornthwaite PE and a soil–moisture storage of 75 mm are given in New Zealand Meteorological Service (1973). These are satisfactory for indicating average seasonal trends, but for detailed 'time and place' comparisons Penman based data are better. Coulter (1975) uses the latter to construct maps of New Zealand showing average number of days per year of 'agricultural drought', average annual 'water deficit', percentage of years with a water deficit, and average water surplus.

Early discussion of irrigation with particular reference to the Waikato may be found in Annett (1953) and Van't Woudt (1956). A later analysis of drought occurrence in the Waikato is that of Baars and Coulter (1974), who, using Coulter's model, predict 20 'drought days' in at least 15 of the 43 years considered. They also found a significant negative correlation between the number of 'deficit days' and pasture production from non–irrigated pasture.

Water Balance Models

Water balance models that have been used in New Zealand differ in their assumptions as to the size and nature of the initial soil moisture store, and the varying rate at which this store is depleted as the soil dries out. They have been designed for specific purposes and there is little doubt that the future will see their continuing development. Of particular value to studies of specific soil–crop combinations will be modifications which model the differential extraction of water from more than one soil layer at a time, at different rates.

To date there have been essentially three models reported. Rickard (1960) used the methods originally suggested by Thornthwaite, and soil moisture changes were calculated until a 'deficit' of 2.04″ (52 mm) was reached, corresponding to permanent wilting percentage in the top 12 inches of Lismore stony silt loam. Later Rickard and Fitzgerald (1969) modified the daily estimate of evapotranspiration to allow for depleting soil moisture. Coulter's (1973b) model uses a soil moisture store the size of which can be varied to suit the application of the model. It assumes that while soil moisture remains in the soil it is withdrawn at potential rates that do not decline as the moisture is depleted. The input is on a daily basis, and the output is on a pentade, decade or monthly basis. It is designed for application to broad–scale climatological studies and is simplified accordingly. Bidwell (1977) describes two models set up for an apple irritation trial: one model has three storages corresponding to three soil layers, and the other has only one storage. Rates of moisture depletion were obtained for the models by fitting actual meteorological and soil moisture data collected over a five year period at the trial site. Their application is discussed by Hewitt (1977).

Lysimetry

Lysimetry provides the most direct solution of the field water balance, and 'drain gauges' and lysimeters of different designs have been used with varying success in New Zealand. The earliest New Zealand lysimeter appears to be that of a Mr Brain of Kaharoa who in 1927 isolated a soil column four feet in diameter (Annett, 1949). Its function was apparently to evaluate phosphate losses, though few results were published. In 1947 a lysimeter was constructed at Ruakura to assess drainage water and nutrient losses (Annett, 1949, 1953). Smith (1953) describes a technique for removing intact cores of 18 inch diameter and two feet length for lysimetry studies. Rickard (1956a) mentions the use of concrete lysimeters (diameter 4 ft, length 3 ft) to determine the amount of excess water draining into the underlying gravels following border–dyke irrigation. The influence of a young stand of radiata pine on the incidence of deep percolation through a free draining pumice soil, and the attendant nutrient losses, are described by Knight and Will (in press) (see also Knight, 1970, 1974a, 1974b) using a collection pad (lysimeter) with 0.3 m side walls and area of 8.5 m^2 located 2.7 m below the surface (Will, 1977; 1966). Rowley (1970) used cut–down 44 gallon drums buried in the ground to measure interception and the net water balance of untreated, burnt, and clipped snow tussocks on the Rock and Pillar Range, Otago. (See also Mark and Rowley, 1969, 1976). Field drainage has been measured by Clothier *et al.* (1977a) using a 2 m^2 $\times$ 1 m lysimeter packed 5 years in advance to replicate layering found in the field. Pearce *et al.* (1977) mentions the use of weighing basket–type lysimeters in forest floor studies. There are no New Zealand reports of the use of weighing lysimeters where the weighing mechanism is an integral part of the system.

SOIL MOISTURE REGIMES

Definition

Moisture regimes define the time variation in the soil moisture status of each horizon within a soil profile, to at least the depth of maximum annual penetration of wetting and drying. An objective of moisture regime studies is to estimate the probability of a particular horizon being within any specified range of moisture content at any time of the year. Such information is especially useful in evaluating the soil factor in irrigation and drainage, and in the assessment of, for example, flood runoff potential, leaching regimes, and optimum conditions for biologic activity.

The moisture regime is thus a statement of the soil storage component of the water balance through time. When measured directly in the field, as opposed to being estimated from the difference between precipitation and evapotranspiration, not only is PE replaced by AE, but any slow drainage within the soil is also monitored. This drainage is of course generally negative, but for some sites can be positive if drainage is contributing water to the site.

Soil Moisture Classification

A soil moisture classification has been defined for New Zealand by Taylor and Pohlen (1962). As currently used, the general concept for each class is that the rooting zone annually meets the following criteria:

Xerous: always drier than wilting point (WP)
Subxerous: drier than WP for 6+ months; always drier than field

	capacity (FC)
Subhygrous:	drier than FC 6+ months and does reach WP; 'Dry' subclass drier than WP 3+ months
Hygrous:	does not reach WP; at or wetter than FC for 8+ months; 'Dry' subclass drier than FC 2–4 months
Subhydrous:	near saturation 6 months; near WP 6 months
Hydrous:	at or wetter than FC all months; near saturation for long periods.

This classification was substantiated by McDonald (1968) using a Thornthwaite water balance on measured available water capacities of the top 18 inches of 54 representative soils. Earlier, Hurst (1951) and Cox (1968) also established correlations of climate and soil groups in New Zealand, and Critchfield (1966) looked at the water–balance approach to climate classification (unfortunately using a very high (300 mm) storage value). The above moisture classes have broad regional relevance, but lack the quantitative definition required by most land use and hydrological studies.

It is pertinent to note that these classes were used in the classification of New Zealand soils at a time when other soil taxonomies preferred accessory properties rather than moisture *per se* (Smith, 1973). There is now a growing trend throughout the world to use moisture data to define taxa; the Canadians and French emphasise the groundwater regime, and the 1975 United States Soil Taxonomy, currently under evaluation in New Zealand, recognises the value of the earlier New Zealand approach by assessing the spectrum of moisture levels. The general concept of each soil moisture class of the United States Soil Taxonomy (U.S.D.A. 1975), with respect to the 'moisture control section' (MCS*) of the profile, is as follows:

Aridic and torric:	– MCS dry in all parts more than half the time (cumulative) that soil temperature at 50 cm is above 5°C – MCS never moist in any part for 90 days (consecutive) when soil temperature at 50 cm is above 8°C
Xeric:	– MCS dry in all parts for 45+ consecutive days after summer solstice, 6 years in 10 – MCS moist in all parts for 45 consecutive days after winter solstice
Ustic:	– MCS dry in some (or all) parts for 90+ days cumulative, most years
Udic:	– MCS not-dry in any part for as long as 90 days cumulative – Perudic subclass: near FC at all times
Aquic:	– reducing regime (free of dissolved oxygen) and saturated – Peraquic subclass: water table near surface at all times.

A comparison of the New Zealand moisture classes and the United States Soil Taxonomy classes is discussed by Watt (1977a) together with comments on the application of the United States classes to the New Zealand situation.

*The MCS is essentially that depth of the profile occupied by the second and third 25 mm depth increment of available water, computing from the surface down; the surface layer is thus excluded.

Regime Results

Field studies in New Zealand have tended to characterise conditions for only short periods or specific seasons, and have often been ancillary to other studies. Notwithstanding, they provide useful information when interpreted in the light of contemporary weather and environment.

Forests

Moisture and temperature regimes are reported by Will (1959) under 7 and 38 year old stands of *Pinus radiata* growing on Rotomahana shallow sandy loam near Rotorua. Data to 3 m were collected approximately every 10 days for a period of one year using fibreglass resistance units. Tensions at 1, 3 and 12 inches (25, 76, 305 mm) are given. At depths below 300 mm the soil did not dry out appreciably even though rainfall was about 17% below normal. In contrast, Jackson (1966) reports wilting point tensions to 0.6 m, and pFs of 3.6 to 3.8 from 0.6 to 1.2 m, in Wingate hill soils (Hutt Valley) under 3 m pine and manuka scrub.

D. S. Jackson (1977) has published seasonal patterns of moisture depletion of yellow–brown sands under young *Pinus radiata* stands of varying density. The study is monitoring the changing moisture regime to a depth of 4–5 m as trees mature at different stand densities. Harris (1974) reports annual moisture regimes in the rooting zone of some coniferous species introduced to New Zealand. Using data from Canada, his main thesis is that vegetation type controls the moisture regime.

Tussock grasslands

In a study primarily concerned with soil creep at Cass (Canterbury), Greenland and Owens (1967) took periodic gravimetric samples at 3 depths to 18 cm on 4 sites over a 10 month period. Day by day changes were estimated by correlating moisture content with rainfall and radiation. Prestidge (1972) also working at Cass reports moisture observations for the period February–April 1972. The top 30 cm of these soils reached wilting point for several weeks in late summer which substantiates their classification as subhygrous. Archer (1976) describes the influence of snow melt on the moisture status of alpine soils on the Ben Ohau Range, and attributes some variability among alpine soils to the period for which they are drier than field capacity.

Mark and Rowley (1969, 1976) collected gravimetric and resistance block data at a site at 920 m on the Rock and Pillar Range in Otago for 28 months. Available moisture never fell below 88% (except when soil froze) and usually exceeded 95%. On the Old Man Range, Central Otago, twice monthly moisture levels in topsoils (0–10 cm) of an altitudinal soil/vegetation sequence have been analysed for the period 1960–62 to indicate possible topsoil moisture regimes of the main members of the sequence. These extend from the brown–grey earths to the high country yellow–brown earths (Watt and Mark, 1973). Similar data for 1960 are reported by Mark (1965a), and regime summaries for Maungatua, Coronet Peak and Old Man Range by Mark (1965b). On the Pisa Range in Central Otago, Wells (1972) reports 0–40 cm gravimetric determinations converted to moisture tensions, for five sites between 490 and 920 m altitude, for the period September 1965 – August 1967. The regime corresponds to the subhygrous moisture class. In the southern Tararua Range Park (1972) checked soil moisture levels on thirteen occasions during the period August 1968 – April 1969 and concludes the gley podzol soils to be rarely at tensions greater than 1 bar. Vegetation/soil drainage relationships of tall tussock communities in the Tararuas are examined by McPherson (1966) and Williams (1975).

Pasture

Routine gravimetric surface (0–10 cm) determinations of non–irrigated Lismore stony silt loam have been collected at Winchmore since the mid 1950's. Some data for topsoils of various soil phases are published for Moutere I.H.D. Experimental Basin (Ministry of Works and Development, 1966, 1967a, 1967b, 1968a, 1968b) but there has been no analysis. Gillingham *et al.* (1974) reports mean monthly topsoil (0–7•5 cm) gravimetric moisture contents for two yellow–brown pumice soils over a three year period, and similar data (Gillingham, 1974) for north and south aspects of a central yellow–brown earth. These studies all contribute to our knowledge of AE from pasture.

Moisture regimes to greater depth under pasture are reported in two studies on yellow–grey earths. Jackson (1966) measured the moisture regime of a yellow–grey earth (Wharekaka silt loam, eastern Wairarapa) for nearly two years. The data imply a regime alternating from extensive drying to the depth of the fragipan in late summer, to very wet conditions through winter. Watt (1977c) gives data for an Otago yellow–grey earth (Otokia silt loam). Observations were made for the period December 1970 to August 1974, and data are presented in terms of the number of days the various horizons are drier than wilting point and wetter than field capacity, thus specifying the results in the same terms as moisture classes. The correlation of such moisture regime data with climate is discussed by Jackson (1976). The significance of moisture to the morphology and performance of yellow–grey earths is also discussed by Pohlen (1956) and Pollok (1975, 1976).

A general qualitative discussion of the hydrologic regime of gley soils is given by Watt (in press).

Site Variability

Where the only water input to a site is precipitation, the estimation of moisture regime from climate data is reasonably straightforward. Most sites however gain or lose water by lateral and vertical drainage and some sites are influenced by groundwater. Such subsurface control on the moisture regime can only be properly evaluated by on–site monitoring.

In the landscape the soil pedologic pattern reflects, among other things, both spatial variation in soil hydraulic properties and site controls on moisture regime. Soil surveys are frequently at too small a scale to show detail of this kind, but slope catena (toposequence) studies have the potential to integrate soil hydrology and pedology. Pedologically oriented toposequence studies of steepland soils in the South Island include Molloy (1974), Archer (1976), and Ives and Cutler (1972). Holloway *et al.* (1962) question whether steepland soils and regolith can be satisfactorily described and classified at all in view of the enormous variability that is sometimes encountered over short distances. Variability of soils on a single steepland slope near Wanganui is examined by Campbell (1973), and the influence of site on moisture properties in soils in the Antarctic is discussed by Campbell and Claridge (1967).

Jackson (1972) has noted the considerable range of variation in soil physical characteristics within catchments at Makara (Wellington). Van't Woudt (1955) describes moisture variation on a volcanic ash soil slope, and Radcliffe (1968) investigates the effect of animal tracking on hill pastures. Hill soil variability in the Hurunui soil set is examined by Tonkin *et al.* (1976). The first studies, however, to correlate slope facets and soil moisture, by relating slope catenas in first–order basins with manifestations of soil moisture variability (if not with the hydraulic properties *per se*), are those of Tonkin *et al.* (1977) and Young *et al.* (1977).

Inter–relationships within small catchments between the soil pattern, the spatial variability of hydraulic properties, and the variability of moisture status, are central to the concept of hydrologic response units. These are areas within which the soil hydraulic characteristics are reasonably uniform. They permit a stratification of the three dimensional topographic and pedologic continuum of a catchment. This can be helpful in model studies of catchment response. Harvey (1974) uses a soil map (1:10,000) to subdivide a high country basin into zones of similar hydrologic response. Murray *et al.* (1975) statistically examine the effect of stratifying a slope sequence on the basis of slope position and slope angle, and claim that such stratification increases the precision of the estimate of mean moisture content of the slope, and reduces the required sample size by half. Budge (1976) in a study of topsoil moisture variability in a small catchment examines the delineation of response zones in terms of slope parameters alone but concludes that soil parameters would improve the definition.

Lawrence (1976) records variations in moisture tension associated with microrelief and dessication cracking in the Wairarapa. Her conclusions are that a cracking soil is generally subject to a greater range of soil moisture conditions than a non–cracking soil, and is generally wetter in winter. An index indicating the rate of change of moisture tension in a profile is shown to be a good predictor of cracking.

Differences in aspect, through differences in both the energy and water balances, can exert a significant effect on soil moisture. This is noted by Archer (1969) and Prestidge (1972) in the Canterbury high country, and by Gillingham (1974), Lambert and Roberts (1976), and Gillingham and Bell (1977) in the North Island hill country.

The interrelationships of site, watertable regime, and drainage properties of soils are brought together in soil survey in the concept of soil drainage classes (Taylor and Pohlen, 1962). Seven classes ranging from 'very poorly drained' to 'excessively drained' describe the duration for which a soil is 'wetter than field capacity, and the rate at which water is removed from the profile'. Drainage classes integrate the concepts of both a permeability classification and a saturation–regime classification, and they have found practical application in soil survey as best described by Cowie (1968) for the Manawatu sand country. Their main disadvantage is that they are only qualitative assessments, and are arrived at using field indicators which may apply only where artificial drainage has not taken place.

SOME APPLIED DIRECTIONS

The foregoing leads to a consideration of the level of our knowledge of soil water in relation to different aspects of land management. It also leads to an appraisal of the extent to which current soil water information is sufficient to make prudent land use and land management decisions in the future.

Irrigation

Irrigation is a major field of management that is concerned with soil water. It must be concluded however that rarely, in the day to day management decision of when to apply water, and how much, is the decision based on any detailed knowledge of the physical properties of the soil, the optimum requirement of the plant, or the current rates of moisture depletion. In rostered border–dyke systems the decision to irrigate depends largely on the availability of water. Sprinkler systems have greater flexibility. With several techniques available for measuring moisture levels and available water capacities in the soil and for predicting with some accuracy the prevailing evapotranspiration rates, and with soil information becoming increasingly available, there is opportunity for refining irrigation practice. Thus water should be applied not by the semi–random approach currently used by most managers, but when the plants most benefit. Increasing costs of energy and water will both help to bring this about.

Opportunities exist too for improving the soil–physical information available to the manager and to the scheme designer. Soil maps can go a great deal further in providing data on infiltration, water storage and permeability, and work is in progress to do this (Watt, 1977d). In providing for improved information, future emphasis must be directed towards:

1. Clearly defining rooting depths. These will, of course, vary with species, but may also be limited by soil restrictions. 'Available' water has little meaning if it is not 'accessible' to roots.
2. Complementing laboratory determinations of the soil moisture characteristic and of hydraulic conductivity with field determinations. The latter are free of disturbance, they characterise natural soil pedons complete with overburden influence, and they reflect the natural continuity brought about by the influences of neighbouring horizons.
3. Developing flow models of water in soil so that complex field situations can be simulated. Thus intermittent and irregular water inputs may be considered entering a heterogeneous soil profile in which the water is redistributing downward while at the same time is being extracted by plants. Such models, based on realistic and meaningful moisture characteristic and hydraulic conductivity data, should go some way toward accurate assessment of water requirements in terms of quantity, timing, soil depth and antecedent conditions.

For further reading on irrigation policy and research in New Zealand the reader is referred to NWASCO (1971), Raeside (1971), Rickard (1973, 1977), DSIR (1974), Griffiths (1975), and Huber (1973).

Drainage

In the design and construction of drainage schemes (Hudson *et al.* 1962; Scott and Mayo, 1972; Bidwell, 1978) soil physical information that meaningfully describes the hydraulic performance of a soil, is essential. Normally however, only subjective qualitative assessments are used to determine such factors as drain spacing, depth, size etc. Improved soil data should be available in the future. This will be essential for the optimum design of more intensive drainage schemes required by more intensive land utilisation. It must be acknowledged, however, that drainage of agricultural lands has been most effective in both islands in overcoming both physical and management problems on soils (Bowler, 1973, 1976). Its value to the agricultural industry has been outstanding.

In drainage, more quantitative information is also desired on the volumes of water involved (e.g. for water harvesting), the effect of drainage on the moisture regime, its modification of streamflow regimes, and its role in nutrient translocation. A lead for such studies is given by Rennes *et al.* (1977), Turner *et al.* (1977) and Cook (1976).

In relation to both drainage and water yield, future studies need to be also directed toward the evaluation of water surpluses in the water balance, the runoff potential of soils (e.g. Heiler, 1977), and runoff process models.

Structure Modification

Considerable research in New Zealand has considered the problem of stock treading and compaction under pasture as related to soil moisture levels (Gradwell, 1956, 1960a, 1965, 1966, 1968b; Brown and Evans, 1973; Edmond, 1974). The practical message from this work is that reduced stocking during very wet conditions greatly reduces the problem. Over–cultivation and/or imprudent cultivation can also lead to compaction problems especially on soils with high silt content. Future research must aim at the accurate prediction of which soils are highly susceptible, and what other parameters are involved in compaction, such as optimum watertable height (Lagocki, 1978). The compaction problem, however, will remain primarily one of management and therefore to be alleviated by good husbandry.

The effect of soil fauna and especially earthworms on pasture production, soil physical properties, and moisture conservation deserves particular attention in the future. Work done to date (Stockdill, 1966) suggests a pronounced effect. The introduction of earthworms to pastures could be, in some circumstances, a sound soil and water conservation practice; evaluation is overdue.

Soil conditioning, through the addition of organic or inorganic substances, is an active field of research overseas. For particular problems it will undoubtedly find application in New Zealand. The general objective (e.g. McDonald, 1962) is to alter the surface structure to give better conditions for plant growth. In some situations it could be to directly improve infiltration, moisture storage, or drainage (e.g. Gradwell, 1958), or to resist erosion. Taken to its ultimate the concept may include work on the constitution of the optimum physical medium for plants, as in potting mixtures (Goh and Haynes, 1977).

The capacity of a soil to act as a medium for effluent disposal is already receiving attention in New Zealand (Wells, 1973; Stevenson, 1976; Childs *et al.* 1977). The field is complex involving hydraulic and other properties associated with the soil's capacity to store and break down loadings. Of vital interest is the extent to which particular plots or fields can cope with sustained loadings over periods of years without the contamination of surface or groundwater resources. The levels of 'soil conditioning' that are attained must be also carefully evaluated.

Environment Assessment

Soil moisture information is an important part of the environmental description, through its clear reflection of climate and soil physical properties, and its integration of site characteristics. In the future, competent ecologic studies will not ignore quantitative soil–moisture description of some form.

Remote sensing of soil moisture by aerial or satellite multispectral photography is a field of exciting potential since at the moment point sampling methods can only provide synoptic coverage of an area at enormous expense.

Both infrared black and white, and infrared colour films may be used for mapping soil moisture variation, or detecting seepage areas (Ross *et al.* 1976; Stevens, 1976; Rijkse, 1977). The reflectance of a wet soil is always lower and therefore appears darker than that of a dry soil. The spectral response of vegetation, however, will in most situations obscure that from the soil, and soil moisture levels may be inferred only indirectly. Thus in aerial photography, crop density and crop vigour may imply moisture differences in a paddock (Ross *et al.* 1976), and well drained and poorly drained areas may be identified via the vegetation's spectral signature. Satellite photography may likewise indicate regional patterns of crop stress or growth, as reported in initial studies in Canterbury (Ellis, 1978). Where ground temperature and moisture levels are related, thermal infrared can be effective provided vegetation is not dense. Passive and active microwave sensors, side–looking radar, and aerial gamma ray survey also offer some potential for soil moisture work, especially through day–and–night/all–weather capability.

The future will also see continuing efforts to model systems involving soil water. Modelling is being increasingly used to evaluate interactions of processes on a conceptual basis. It is complementary to laboratory and field studies and achieves its most useful application when continually modified and updated in accord with the findings of contemporary research on constituent parameters. Examples include Fitzharris's (1974) evapotranspiration model as applied to a hypothetical land use comparison of the water balance, and the simulation model of a grazing system (Wright and Baars, 1976; Baars *et al.* 1977) where soil moisture and pasture production components are combined to determine irrigation strategies and the most profitable allocation of water. Catchment models attempt to simulate the modifying influence of vegetation, soil, and regolith on the rainfall/runoff relationship (Boughton, 1968; Hutchinson and Simmers, 1971; Murray, 1970; Wood and Sutherland, 1970).

CONCLUSION

What do we know about water in New Zealand soils? This review has been widely based to answer this question, not from one scientific point of view, but by considering the application to physical hydrology of several kinds of soil water studies. As such, this catalogue of the New Zealand experience brings together a considerable literature which is currently escalating as the result of a number of factors. These include: growing appreciation of soil water as an important ecologic factor in environmental research; growing appreciation of the role of soil in watershed performance; increasing need to economically optimise agricultural systems and to design irrigation and drainage schemes from a scientific base; a response to an upsurge in international interest in soil physics and soil hydraulics with improved methods of measurement; and a response to the expansion and development of soil physics theory.

From this it may be concluded that soil-water is a dynamic field of research which is only just beginning to evaluate of the role of soil and soil water processes as first order modifiers of the precipitation/runoff/groundwater relationship. The future promises considerable development in both theoretical and applied directions.

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GROUNDWATER – STATE OF THE ART IN NEW ZEALAND

H. R. Thorpe and D. M. Scott

ABSTRACT

This is a general review of groundwater investigation, research and management problems in New Zealand by universities, government departments and regional water boards. It outlines the methods currently in use and the results being achieved in field investigation, data acquisition and management and the applications of computers to date in data analysis and groundwater modelling. The problems of ground water quality which are being encountered in New Zealand are mentioned.

There is an extensive bibliography of recent work published in this country.

INTRODUCTION

The abundance of available surface water in New Zealand has reduced the need to develop existing groundwater resources. Partly because of this the state of the art of groundwater hydrology is less advanced than most other aspects of hydrology. However, population increases and economic pressures have brought about the realisation that sooner or later all our water resources must be efficiently developed. Even now in some areas of urban development such as the Hutt Valley or rural areas of intensive agriculture such as the Waimea and Heretaunga Plains problems of resource management are present.

The management of the nation's water resources is the responsibility of the regional water boards. A survey of the boards indicates that the management of underground water is taken seriously although the level of activity varies widely, reflecting the different pressures which exist in each region.

Other organisations, which are involved with groundwater research rather than management, such as the research and survey group in the Water and Soil Division of the Ministry of Works and Development, and the New Zealand Geological Survey of the Department of Scientific and Industrial Research, provide specialist assistance to the boards when requested. Particular skills also exist within the universities and other divisions of DSIR such as Chemistry Division (water quality), Nuclear Sciences (isotope tracers), Geophysics (seismic and resistivity surveys) and Physics and Engineering (numerical modelling). The Cawthron Institute also has experience in water quality aspects of groundwater investigation and may be used on a consulting basis.

Management Problems

The major problem facing regional water boards in connection with groundwater is that of the equitable management of a limited resource to meet existing and future demands. With one or two notable exceptions the knowledge of groundwater systems in New Zealand is quite inadequate for management purposes thus making the control of water rights difficult.

Other problems relate to the protection of water quality from the effects of unwise land use including pollution of coastal sand aquifers by septic tanks at resort areas. Salt water intrusion in coastal aquifers caused by overdrawing is also a concern in some areas.

DATA ACQUISITION AND STORAGE

As in other fields of hydrology, an increasing programme of groundwater investigation is creating a large pool of data which must be processed and available for efficient use. There are two basic types of data: time–series such as water level or water quality records; and stationary (non–time series) data such as well logs and hydraulic characteristics. To date the bulk of data gathered has been the stationary type, and its collection and storage unco–ordinated. Each type of data requires different processing and two distinct systems have been developed in New Zealand.

TIDEDA, the time dependent data handling system, which has been developed by the Ministry of Works and Development includes facilities for filing ground water data such as instantaneous water levels or water quality parameters (Ibbitt and McConnell, 1974). Up to 15 water quality parameters encompassing physical, chemical or bacteriological data can be recorded for any well. The ability to store water quality parameters has not yet received much use. Although TIDEDA is designed for time–series data, stationary data such as identification and hydraulic details, aquifer lithology and well construction can be stored. The TIDEDA system was used by Donaldson and Campbell (1977) to assist in the study of the Hutt Valley aquifer, and it is receiving growing use by other workers.

A data storage system which is in many ways complimentary to TIDEDA has been developed by the Geological Survey and forms the basis of a standard form for the recording of stationary water well data (Brown and McCammon, 1976). Though the computer system has yet to receive significant use the standard form is being used and some old records have been transferred to it. The system is of the computer based information system type and makes use of an interactive man–computer dialogue to provide access to filed data.

Some regional water boards require the holder of a water right for groundwater extraction to report on usage, as a condition of the right. This will be valuable in the long run if stored as a time series as it can then be matched with piezometric fluctuations.

INVESTIGATION METHODS

The basic objective of an investigation of groundwater resources is to establish the quantity and quality of the available water. The investigation methods involved can be divided into hydraulic and geological methods.

Hydraulic Investigation

The hydraulic investigation is intended to provide data which describe the hydraulic parameters of the aquifer; such as the distribution of transmissivity. Several types of data are normally required: piezometric surface levels, measured or estimated flow directions and velocities and response to aquifer pump tests. Where dispersion of pollutants is of concern it may also be necessary to measure the dispersivity of the aquifer.

The basic method of obtaining piezometric levels is by direct observation in bores. While this is a routine matter where wells exist, considerable effort is required to locate and level an adequate network from which contours and flow directions can be derived. Though piezometric levels may change slowly, groundwater systems are dynamic and hence continuous records at a few selected points are necessary in acquiring an understanding of the response of the system to various inputs and withdrawals. To date surprisingly few such records exist in New Zealand. Piezometric levels have also been measured using geophysical methods (Risk, 1967, 1974; Borgesius, 1975) and more use could be made of such techniques with advantage.

Regular piezometric surveys are of assistance in determining long term trends of the piezometric heads and locating aquifer boundaries. Piezometric contour maps are also useful for the calibration and verification of numerical models of groundwater flow and it is possible to use the maps to determine the distribution of aquifer transmissivity. Hunt and Wilson (1974) have described a graphical technique to calculate transmissivities and steady–state flow rates for groundwater flow in northern Canterbury.

The standard method of determining the aquifer characteristics of transmissivity and storage coefficient is the aquifer pump test. To obtain reliable information from such a test it is necessary to have a pumped well with several observation wells at various distances. As this can be an expensive requirement pump tests are only done for special or large scale investigations. Methods of analysis are found in standard geohydrology texts. To date most of the limited New Zealand experience has been gained in the Heretaunga and Canterbury Plains (various unpublished reports of Ministry of Works and Development, Napier District, North Canterbury Catchment Board, South Canterbury Catchment Board).

Though of assistance in establishing aquifer parameters for resource evaluation the measurement of flow direction and velocities has particular relevance to the study of contaminant movement within an aquifer. Martin and Noonan (1977) and Ministry of Works and Development (1977) have established that groundwater flow directions in the scale of a few hundred metres can differ substantially from directions deduced from large scale piezometric maps. Techniques now being used to directly determine the direction and velocity of groundwater flow employ a variety of tracers in either single– or multiple–well tests.

Work on groundwater velocities with tracers appears to have begun with the studies on tritium in the Heretaunga Plains (Grant–Taylor and Taylor, 1967) and has expanded from the use of natural isotopes to the use of a variety of injected tracers. Recent experience (McCabe and Rowse, 1976) has shown that the radioactive tracer iodine–131 does not function well in soils, as it appears to adsorb on organic matter. It has, however, been used more successfully in gravels below the soil. Thorpe (1977) reports on the successful use of lithium chloride, bromine–82 and the fluorescent dye rhodamine WT in studies of the gravel aquifers of the Heretaunga Plains. Pyle and Davis (1977) have at the same site made very successful use of a micro-organism as a tracer. *Escherichia coli* (H_2S+) is a non–pathogenic mutant strain which, when cultured, produces colonies with a black centre, easily distinguished from other soil micro-organisms. *E. coli* (H_2S+) travelled freely in these gravels with velocities which were consistently slightly *greater* than those for chemical tracers. The micro-organism was detected in substantial numbers at a distance of 125 metres from the point of injection. Die-off experiments both in the aquifer and in a dialysis sac suspended down a bore gave T_{50} of 2.2 days with micro-organisms still being found in small numbers at 26 days.

Chloride is one of the most conservative ions. As sodium chloride, it is cheap and readily available, but cannot be used as a tracer if background levels are high. Where background levels are low, large volumes (20 m^3) of injected sea water have been successfully detected at distances up to 300 metres either directly with a conductivity meter or specific ion electrode, or indirectly (to 125 metres) by resistivity survey (Thorpe, 1977).

The latter method has only been used where the water table is within 12 m of the surface, where velocities are greater than 4 m/day, and where the surface formation is of reasonably high resistivity, as in dry gravels or sands, to allow deep penetration of the current. Under these conditions the resistivity technique will give ground water flow direction and velocity more quickly and cheaply than by using standard multi-bore tracer techniques.

Less exact methods have been tried (McCabe, pers. comm.) involving the injection into a screened well of a radioisotope with a short half-life. If a non-adsorbing isotope is selected, the rate of reduction of radio-activity in the well gives a measure of the groundwater velocity. If an adsorbing isotope is used it will remain on the aquifer material near the well and by suspending a collimated detector down the bore on a rod and slowly rotating it the position of the adsorbed material and hence the approximate direction of flow can be established. These methods are more fully described by Drost (1974).

The selection of a groundwater tracer depends upon the problem under consideration; and availability, cost, lower limit of detection, toxicity and ease of analysis of the tracer. For many situations rhodamine WT would be the most suitable.

On a larger scale, origin of groundwater can sometimes be determined by stable isotope studies using tritium, deuterium and oxygen-18. This has been done for the Rotorua region, where several springs were examined, and it was possible to distinguish between springs which are fed from the lakes and those which are not (Taylor, Freestone and Nairn, 1977).

When considering groundwater pollution where a knowledge of pollutant behaviour within an aquifer is required, it may be necessary to consider the dispersivity of the aquifer. The direct measurement of dispersity is a very difficult process involving an array of wells so that when the tracer is injected at a point an indication of the movement in three dimensions can be obtained. Values have been computed for the gravel aquifers of the Heretaunga Plains in a recent study (Ministry of Works and Development, 1977) and are shown in Table 1. These values are mostly from measurements made over a flow path of up to 77 metres. Also included are values obtained from measurement of chloride ion movement over a distance of about 300 metres in leachate from a nearby rubbish pit but these results, computed from data gathered for other purposes, must be considered less reliable (Ministry of Works and Development, 1977).

The range of values shown are an indication of the heterogeneous nature of these gravel aquifers which have been laid down by a braided river. In some cases the distribution of tracer within a well did not fit the dispersion theory being used, and values could not be computed.

The velocities measured in these experiments are exceptionally high by world standards but are consistent with values obtained in Canterbury gravels (McCabe and Rowse, 1976).

TABLE I – Dispersivities and mean velocities calculated from experiments on the Heretaunga Plains.

Site	α_{xx} *(m)*	α_{yy} *(m)*	α_{zz} *(m)*	*Mean velocity (m/day)*
Roys Hill	1.4-11.5	0.1-3.3	0.04-0.1	11.4-230
Flaxmere	0.3-1.5	Not measurable	0.06	10.2-90
Hastings City Rubbish Dump	41	10	0.07	6.4

Where α_{xx}, α_{yy}, α_{zz} are the longitudinal, lateral and vertical dispersivities respectively.

Geological Investigation

The basic method of acquiring knowledge of the geology of an aquifer is based on the carefully logged borehole. Such logs, in their thousands, are held at offices of the Geological Survey as well as by regional water boards and the Ministry of Works and Development.

Within the Geological Survey, most of the experience is centred in its Christchurch office and attention has been focussed primarily on the Canterbury aquifers. Wilson (1972, a, b; 1973, 1976) has summarised this experience from a broadly geological viewpoint. Unfortunately the most economically significant aquifers in the country are those composed of quaternary gravels which are very difficult to sample and log in a manner which accounts for known hydraulic variations. Furthermore these logs are usually obtained as a by–product from bores sunk for purposes other than geological information and their quality is undoubtedly variable.

Geophysical methods can be used to extrapolate from limited borehole information to give greater areal coverage, and also to derive more information from each bore. Some of these methods have been used successfully in New Zealand already but on a limited scale.

Hochstein and Lawton (1976) have employed gravity and magnetic measurements in the Auckland metropolitan area to delineate the ancient topography underlying the later volcanic deposits.

Risk (1967, 1974) has used resistivity techniques in both the Canterbury and Heretaunga Plains to determine the depths to water table and thicknesses of overlying strata. Borgesius (1975) has employed this method to map basement beneath the unconfined aquifer of the Heretaunga Plains to depths of 250 metres. Seismic surveys are known to be useful for locating the water table in gravel aquifers but have been used to only a minor extent for this purpose

(Ingham, 1963).

Although used extensively overseas in groundwater investigations, downhole geophysical techniques have not yet been used productively in New Zealand.

These techniques can be used to define lithology by measuring such properties as: electrical resistivity, emission of natural gamma radiation, electrical potential differences at boundaries of strata and back scattering of gamma or neutron radiation. From this information qualitative information on density, porosity, clay content and hence permeability can be derived. With experience and careful calibration in a particular area these methods can produce quantitative data (Keys and MacCary, 1972). Basic equipment is now held by the Ministry of Works and Development and is being evaluated under New Zealand conditions.

Water Quality Investigation

Groundwater quality can be affected by the nature of the water at the point where it enters the ground and by the geochemistry of the material which forms the aquifer. Thus in addition to isotopic techniques the chemical nature of the water can on occasion be used to deduce information about water origins and movement. Various anions and cations when considered in combination impart a 'signature' to the water and when related to details of the aquifer geology act as natural tracers. The methods have been widely used overseas (Stiff, 1951; Piper, 1953) and an example of their use in New Zealand is contained in a report by the Auckland Regional Water Board (1977). Usually when sampling groundwater, the investigator is forced to make use of existing wells and this requires a period of pumping before sampling to ensure that the bore is thoroughly flushed. The sample is therefore representative of a large volume. This approach can be misleading unless consideration is given to the structure of the aquifer and the depth of the well.

There is evidence that where pollution is suspected it may be more severe close to the water table, (Saffigna, 1977; Biegler, 1961), therefore where public water supplies are to be monitored more detailed studies are warranted. This would involve taking grab samples through a long screen in a well, or having a series of sampling wells set at various depths. One such study is underway in the Ngatarawa Valley near Hastings at the present time (Thorpe, pers. comm.).

GROUNDWATER QUALITY AND LAND USE

A significant number of the deeper aquifer systems in New Zealand contain non–potable water which has been mineralised, or is corrosive. In addition it is now realised that the natural groundwater quality can be significantly affected by the pattern of land use above an unconfined aquifer.

Fully serviced residential urban development does not appear to pose a serious risk to groundwater quality. The Hutt Valley aquifer, after thirty years of urbanisation of the valley floor, still produces water of high quality. On the other hand, in situations where water supply is from private wells and sewage disposal is by septic tank, problems have occurred. This has been observed in coastal holiday resort areas and in low density residential areas on the outskirts of urban communities.

Similarly urban waste disposal on land, whether solid or liquid, has a high potential for pollution and at least one case has been documented in New

Zealand (Ministry of Works and Development, 1974). A groundwater study should be an integral part of any investigation of waste disposal sites.

Potentially much more serious is the effect of intensive agricultural activity, especially if it is combined with irrigation which causes increased leaching of soils. The most serious effect apparent in groundwater beneath intensively farmed land in many areas in New Zealand is the increase in nitrate levels (Martin and Noonan, 1977; Saffigna, 1977; Ministry of Works and Development, 1977). As nitrate is regarded as potentially toxic to infants at quite low concentrations and as it is not easy to remove nitrate from water supplies, this is a matter of some concern.

A plant nutrient commonly applied to agricultural land is phosphorous in superphosphate. However phosphorous compounds generally bind tightly to soil particles and do not penetrate more than a few centimetres below the surface, hence phosphorous levels in ground water are low.

Although little work has been done in New Zealand on the presence of biocides in ground water, overseas experience strongly suggests that because of the affinity of most of these compounds for soil particles, and/or their rapidity of breakdown, they will not appear in ground water in significant quantities (Croll, 1972; Steenvoorden, 1976).

COMPUTER TECHNIQUES IN INVESTIGATION AND MANAGEMENT

The modern developments of high speed electronic computers with large memory have provided a valuable tool in the study of groundwater. The areas of principal impact have been in the numerical modelling of regional groundwater flow and aspects of data storage and manipulation – though there are many computer applications in groundwater studies that fall outside these two fields.

Groundwater Modelling

Though modelling of ground water flow can be performed in a number of different ways, in the majority of cases computer–based numerical methods are used to solve a system of partial differential equations, boundary and initial conditions. Prickett (1975), in discussing ground water modelling identifies analog models (e.g. Hele–Shaw models, resistance–capacitance networks); analytical and numerical methods (e.g. Theis equation, numerical analysis by finite difference methods); and sand tank models.

Because the high speed large memory computers required for numerical methods are now available in New Zealand it seems likely that numerical models will be used in the majority of situations where regional modelling of ground water flow is necessary.

Quantity models

In New Zealand there are two geographical areas where regional modelling of groundwater behaviour has been carried out in some detail. Analysis of aquifers in the North Canterbury Catchment Board's area has been carried out by staff and students of the School of Engineering, University of Canterbury in co-operation with the Board. The Hutt Valley aquifer has been studied by staff of the Physics and Engineering Laboratory of DSIR.

In the first of a series of reports on the aquifer of North Canterbury, Hunt (1976a) has presented the results of a groundwater simulation study for the area between the Selwyn and Rakaia Rivers. In this study, a flow net analysis gave a transmissivity distribution which then was verified with a numerical simulation. The calculated transmissivities were used to construct both steady and unsteady seepage models, and several examples illustrated the use of the numerical models for predictive purposes. Finite difference methods were used.

Day (1976) has applied the same approach to develop a steady–state model for the region between the Selwyn and Waimakariri Rivers. The characteristics of the area required a knowledge of the distribution and magnitude of major well flow, and leakage losses from the aquifer.

Hunt (1976b) described two new computer models for groundwater flow. These were improvements of the steady–state and unsteady–state models described in his earlier report. The application of both models was demonstrated by applying them to the Selwyn–Waimakariri area using the finite difference mesh, transmissivities and leakage coefficients calculated by Day.

Richards (1976) has developed Hunt's later unsteady model to give better estimates of drawdown near pumped wells. His procedure works in conjunction with a regional model to provide the boundary conditions for a smaller scale, higher resolution model centred at the pumped well.

Donaldson (1974) developed a two dimensional time–dependent computer based model of the Lower Hutt – Port Nicholson groundwater basin. The model simulates the ground water system under real operating conditions as well as predicting effects of alternative development proposals. Finite difference methods were used with quadrilateral elements rather than the usual rectangular grid. Donaldson has based his finite difference equations on the water balance for the individual elements.

Donaldson and Campbell (1977) extended the above analysis to compare the finite difference method and the finite element method. In a comprehensive report of the experience gained in simulating the behaviour of the ground water basin, they conclude that there is no significant difference between the results obtained with the two methods and chose to use the finite element method as it used less computer time. In this application the finite difference method was based on a rectangular grid with a variable spacing. For the finite element model the Galerkin residual technique was used to formulate the equations and an algorithm developed by Pinder and Frind (1972) was followed. Donaldson and Campbell used the finite element model to perform an analysis of the hydraulic effects of present and future use of the water resource of the basin. This analysis showed limits to the system and the effects of well placement and management.

Quality models

Water quality models, like quantity models, can be based on analytical solutions to idealized problems or numerical solutions to problems with complex characteristics. Though the analytical solutions have many limitations, in the absence of sufficient data to use numerical solutions sensibly, the analysis of an idealized problem can provide useful information about the behaviour of a mass–transport system.

This approach was used in a study of contaminant dispersion in the Heretaunga Plains aquifer (Ministry of Works and Development, 1977). That study made use of a computer based simulation of contaminant dispersion within the saturated zone of the aquifer using a steady–state two part analytical model of dispersion. Though the simulation made use of simplifying assumptions about the hydraulic characteristics of the aquifer it provided a useful basis for discussion. The model demonstrated that pollution levels, from an extensive area such as a residential development, are insensitive to longitudinal and lateral dispersion rates. The vertical dispersion rate and vertical configuration of the aquifer were identified as being the most significant factors in the distribution of pollutants within the aquifer.

Computer Aided Contouring

Hunt (pers. comm.) has developed a computer programme to assist in the preparation of piezometric contour maps. Using a least–squares technique the programme treats sub–areas of the region by fitting a three dimensional surface to the observed water levels. The fitted surface is used to generate the piezometric contours and streamlines and also allows the calculation of relative transmissivities. The procedure was developed for use by the North Canterbury Catchment Board in the area between the Rakaia and Waimakariri Rivers. The programme eliminates some of the drudgery of manual contouring and provides a valuable objectivity to the process so that contour maps produced from different piezometric surveys can be compared.

The Identification Problem

The identification problem involves the inverse of the problem of modelling ground water flow. Instead of predicting ground water levels from the transmissivity and storage coefficient, the inverse problem involves the determination of these parameters from observed ground water levels. Shiati (1977) has developed a mathematical model which, using a computer programme to fit a second–degree polynomial surface to observed piezometric heads, allows the identification of the ratio of storage coefficient to transmissivity. Spatial and time derivatives of this fitted surface are used to derive the ratio of storage coefficient to transmissivity. The method was tested on field measurements made over an area of several hundred hectares near Lincoln, in Canterbury.

Time Series Analysis

McCammon (1976) has used computer techniques in a study of long term well hydrograph records in Christchurch. In an analysis of the seasonal pattern of change in static level on an areal basis he used computer generated plots, contour maps and correlation diagrams to facilitate interpretation, and to demonstrate that static levels have maintained a steady–state condition despite an increase in population and increased water demand.

Using a similar approach, Donaldson (1977) analysed the natural variations in well heads in the Waimakariri–Rakaia region of the Canterbury Plains to determine the background flow pattern of the system and the predominant source of the water in its various sections. Studying the correlations between well levels in various wells, between well levels and rainfall infiltration, and between well levels and river levels, he concluded that rainfall played a greater role in the recharge of groundwater than water infiltrating from rivers. This conclusion is a reversal of the widely held view that rainfall plays an insignificant or minor part in groundwater recharge.

OTHER ASPECTS OF GROUNDWATER RESEARCH

The emphasis in this paper has been upon the more orthodox type of aquifer system. There are however two other lines of research underway in New Zealand which are noteworthy: geothermal groundwater and karst aquifers.

A substantial effort has been made for many years to investigate geothermal groundwater resources and with the present (1978) energy crisis this effort is likely to expand considerably. The problems faced in geothermal investigation are broadly similar to those facing the orthodox geohydrologist but with the added complications of working at greater depth, temperature and pressure.

The two fields which have been most intensively investigated are those at Wairakei and Broadlands, with some interest being shown in developing the Kawerau field to supply the Tasman Pulp and Paper Company (Bolton and Studt, 1977). Attempts have been made to model the Wairakei field and although some success has been achieved, it is still not possible to predict field behaviour under different development schemes with confidence (R. S. Bolton pers. comm.).

Interest in karst hydrology in New Zealand is centred in the Geography Department, University of Auckland. The aquifers on which most work has been done are in the Waitomo area of the North Island and the Takaka district of the South Island. Results of this research to date are incorporated in papers by Williams (1977) and Gunn (1977).

CONCLUSION

Groundwater hydrology is a field of endeavour which requires a multi–disciplinary approach, incorporating the expertise of the geologist, physicist, chemist, hydraulic engineer, mathematician and water resource manager.

The objective of most groundwater research in New Zealand at the moment is to gain a sufficient understanding of the aquifer systems to enable soundly based management. This research may be at the level of developing techniques of investigation, whether in the field or the computer room, or at the point of producing computer based management aids for regional water boards. Generally, although not exclusively, the universities direct their work towards the more basic and detailed studies while the Government departments are involved with broader scale problems, and this complimentary approach is to be encouraged.

While information of any type relating to aquifer characteristics is an aid to management, it is highly desirable that some tool enabling *prediction* of aquifer response under various management systems or assumed trends in water demand should be available. Water demands on many of the aquifers are increasing rapidly and it will be necessary to have management policy formulated and understood by the water users before competition for the resource becomes more intense. The ultimate objective of groundwater research should be the development of models to allow prediction of aquifer behaviour. Fortunately it is not necessary to have the complete body of information before a model can be constructed. With simple information, simple models can be built which not only provide management information but guide further investigation. As more information is gathered, models can be rebuilt or refined to improve predictive capability.

On the assumption that groundwater demand will continue to rise, together with the need for the resource to be managed, it can be safely predicted that the next few years will see a rapid development in the whole field of groundwater research, from basic data collection to the production of sophisticated resource management aids.

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EVAPORATION FROM LAND SURFACES

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ABSTRACT

The physical theory of evaporation from soil, plant leaves and vegetated surfaces is briefly reviewed. Against this background, the physical basis of various evaporation estimation procedures is discussed. It is observed that physical models are seldom used in practice since sufficient data sets are usually unobtainable, and that empirical methods are substituted. Simple empirical methods using radiation data can give satisfactory results when strong advective effects are absent, but their reliability can be greatly improved by local calibration. When advective enhancement is strong local calibration is particularly important. The discussion is illustrated with some previously unpublished evaporation data from Palmerston North.

INTRODUCTION

Evaporation is the process whereby water is transformed from the liquid to the gaseous state. This phase change may occur at the soil surface, at the surface of water droplets held on plant parts following rainfall, dewfall or irrigation, and at the moist cell walls within plant leaves. In this last case the water vapour must escape to the atmosphere through the numerous small stomatal openings in plant leaves through the process known as transpiration.

Making accurate field estimates of evaporation is a very practical concern. For much of New Zealand's productive agricultural and forestry regions a large part of the rainfall is returned to the atmosphere by evaporation. Soil water deficits are common in late spring and summer, and where irrigation systems are not used to maintain high rates of production at these times, dryland management practices must be adopted. Runoff into the rivers is sensitive to evaporation rates which to some extent can be altered by land management regime. The river flows depend on land use pattern in the mountain and hill catchments.

Evaporation estimation accuracy must be consistent with the use to which the data are put. In regional water resource surveys involving large catchments and the planning of irrigation schemes, monthly data with 15–25% accuracy can be satisfactory. Water balance studies of crops, pastures and forests, on-farm irrigation schedules and estimates of water reservoir losses may demand daily estimates of 10–15% accuracy. For intensive studies of the soil-plant-atmosphere system with hourly measurements, accuracies approaching 5% may be preferred (Tanner, 1967).

There are many good reviews of evaporation (e.g. chapters 26-29 in Hagan *et al.* 1967). For a detailed treatment of the micrometeorological theory of evaporation and field measurement of evaporation rates, the reader is referred to these or other works. Here the attempt is to review the state of knowledge of evaporation and to present the authors' own overview and perspective. The physical basis and achievements of various models of evaporation from plant and land surfaces are discussed, without attempting to detail the mechanics of those models.

A more practical discussion of the problem of evaporation estimation follows. By setting these estimates against the theoretical background, the intention is to give a clearer understanding of the physical basis and limitations of the various methods. Where possible local examples are used to illustrate the arguments.

Evaporation from the Soil Surface

When the soil surface dries the evaporation rate remains identical to that from a saturated surface for some time. During this period evaporation is determined by the prevailing meteorological conditions.

Subsequently the ability of the soil to conduct liquid water to the surface becomes a limiting factor. At this time evaporation is restricted by soil physical properties. Finally, vapour phase transport predominates resulting in very low evaporation rates of little practical importance. This separation of soil drying into three phases was developed by Philip (1957), and has been identified in the field (Idso et al., 1970). From a water–budget modelling point of view, this separation is very convenient.

In the second phase of drying, modelling evaporation amounts to finding solutions to the non-linear flow equation:

$$\delta\theta / \delta t = \delta / \delta z \, [D(\theta) \ \delta\theta / \delta z] \tag{1}$$

where $D(\theta)$ is the hydraulic diffusivity of the particular soil. Equation (1) assumes an isothermal soil and ignores gravity but these simplifications lead to small errors (Philip, 1957; Hanks *et al.*, 1967).

Complete solutions to (1) are difficult and often impractical. A simplified solution has been developed by Gardner (1959) viz:

$$E_S = mt^{-½} \tag{2}$$

where t is time from the start of the drying cycle and the proportionality constant depends on the weighted mean hydraulic diffusivity of the soil. This solution has been used successfully by Black *et al.* (1969) for Plainfield sand and Kerr (1974) for Manawatu fine sandy loam. In modelling daily soil evaporation, Black *et al.* (1970) calculated both a meteorological estimate assuming that the soil surface was wet, and a second soil–limited estimate from (2). The smaller value was used in their water budget calculation. Tanner and Jury (1976) have adopted a similar basic procedure though the two estimates were calculated somewhat differently.

This procedure of estimating E_S is the most satisfactory available at present and can be used to obtain estimates for the crop water-use models developed in later sections.

Micrometeorological Transport Processes

As part of the process of evaporation, water is transported away from the various leaf and soil surfaces into the turbulent air-stream above. In the models which follow, this process is described in terms of various transport resistances. The purpose of this section is to very briefly introduce and explain these resistances. A much fuller account is given by Thom (1976).

In molecular diffusion of gases, small scale random movements of molecules cause a net migration from regions of higher concentration to regions of lower concentration. Smaller, more mobile molecules diffuse more rapidly, in inverse relationship with the square root of their molecular weight. An equation describing water vapour diffusing through the laminar boundary layer surrounding a leaf can be written as

$$E = -\kappa_E \ dq/dz \tag{3}$$

where q is the water vapour concentration or absolute humidity, κ_E is the molecular diffusivity for water vapour and E is the evaporation rate.

In turbulent transport, larger scale random motions of whole parcels of air transport heat and vapour much more efficiently. Within a few tens of meters of the ground the fluxes are directed down concentration gradients and transport is often described, by analogy with (3), using 'eddy diffusion' equations. In the case of water vapour this can be written as

$$E = -K_E \, dq/dz \tag{4}$$

where K_E is the "effective" or "eddy" diffusivity for water vapour. In contrast to κ_E, K_E does not depend on the local properties of water vapour, but is a property of the flow, notably the size and time scales of the turbulent motions.

If the water vapour flux is constant with height, then (4) can be integrated from the surface up to some height z to give a finite difference form

$$E = (q_o - q_z)/ r_{aE} \tag{5}$$

where r_{aE} is the aerodynamic resistance to vapour transport. It is common practice to express humidity as a vapour pressure (e) in millibars rather than a concentration (1 mb = 100 Pascals). Expressing the water flux as an equivalent latent heat flux, (5) can be rewritten as

$$\lambda E = \rho C_p \, (e_o - e_z)/ \gamma r_{aE} \tag{6}$$

Similar expressions can be written for sensible heat and momentum transport. Thus:

$$C = \rho C_p \, (T_o - T_z)/r_{aC} \tag{7}$$

and

$$\tau = \rho u/r_{a\tau} \tag{8}$$

respectively. The friction velocity, u_*, is defined by

$$u_*^2 = \tau/\rho \tag{9}$$

so that $r_{a\tau}$ is given by:

$$r_{a\tau} = u/u_*^2 \tag{10}$$

The friction velocity can be found from micrometeorological measurements.

It is often assumed that the effective diffusivities for any transported species are equal. The weight of experimental evidence supports the assumption that the effective diffusivities for sensible heat and vapour are equal ($K_C = K_E$) (Swinbank and Dyer, 1967), but not that $K_C = K_\tau$ (Businger *et al.*, 1971; Pruitt *et al.*, 1973) where K_τ and K_C are the effective diffusivities for momentum, and sensible heat, respectively.

For heat and water vapour transport Thom (1972) argues that transport resistance should be calculated from:

$$r_{aC} = r_{aE} = r_{a\tau} + 6{\cdot}3 \, u_*^{-2/3} \tag{11}$$

Evaporation from a Single Leaf

A basic element of any plant canopy is a single leaf. It is helpful to briefly consider the evaporation from a single leaf and how this can be modelled.

A typical leaf is thin enough that temperature gradients across it are very small and can be ignored. The atmosphere in the intercellular spaces within the leaf can be assumed to be at the saturation vapour pressure corresponding to leaf surface temperature (e_o^*). This is because water potentials in viable leaves are rarely below –40 bars, and at that potential the equilibrium relative humidity is still 97% of saturation.

Stomata comprise a diffusive resistance to the outward flow of water vapour through the leaf surface, and transpiration can be described by the equation:

$$\lambda E = \rho C_p \, (e_o^* - e_o / \gamma r_s \qquad (12)$$

where r_s is the stomatal resistance for unit surface area of leaf. Equation (12) is not particularly useful by itself because surface temperature and humidity are not easily measured. Fortunately, evaporation from the leaf can be related to more easily measured parameters by following the procedure of Monteith (1965).

Consider a single leaf surface with a net radiant energy income per unit area of R_n. Some of this energy may be conducted through the leaf (G), but the rest is dissipated into the airstream over the leaf as either sensible or latent heat. Symbolically, this statement of conservation of energy can be written as:

$$R_n - G = C + \lambda E \qquad (13)$$

The quantity ($R_n - G$) is called the available energy.

Along the surface of a leaf exposed to the wind a thin layer forms within which velocities are reduced by viscous drag forces. Convective transport in this layer is much less efficient than in the turbulent airstream above and is usually modelled by the device of a 'boundary-layer' resistance so that equations of the form of (6) and (7) can be written. Thus:

$$C = \rho C_p (T_o - T_z) / r_{bC} \qquad (14)$$

and

$$E = \rho C_p (e_o - e_z) / \gamma r_{bE} \qquad (15)$$

where r_{bC} and r_{bE} are the boundary-layer resistances to sensible heat and vapour transport respectively.

Both T_o and e_o may vary down the surface of the leaf and the mean values are used in (14) and (15). The resistances r_{bE} and r_{bC} can be related to the flow velocity, leaf dimensions and molecular properties of the diffusing species. Usually it is adequate to assume $r_{bE} = r_{bC}$. Murphy and Knoerr (1977) describe a recent determination of an equation for calculating boundary–layer resistance, together with a summary of previous work.

Algebraic manipulation of equations (12) through (15), together with the definition:

$$s = \left. \frac{de^*}{dT} \right|_{T = Ta} \approx \frac{e_o^* - e_z}{T_o - T_z} \qquad (16)$$

then gives

$$\lambda E = \frac{s(R_n - G) + \rho C_p \, (e_z^* - e_z)/r_b}{s + \gamma (r_s + r_b)/r_b} \qquad (17)$$

which is the equation developed by Monteith (1965). If the leaf is completely symmetrical then total evaporation from both sides of the leaf is expressed by (17) with the reinterpretation of R_n as the sum of the net radiation into both sides of the leaf, and G set to zero since energy conducted into one leaf surface must flow outwards at the other.

The thermal radiation emitted by the leaf depends on leaf surface temperature. Equation (17), as it stands, cannot be used predictively unless the final surface temperature can be predicted. The model can be extended to allow for changes in outward thermal radiation, but that is outside the scope of this brief discussion.

Evaporation from Plant Canopies

Evaporation from a whole canopy is the sum of evaporation from the leaves plus soil evaporation. Thus:

$$\lambda E \;=\; (\rho C_p/\gamma \hat{A})\; \sum_i A_i(e^*_{oi}-e_{oi})/r_{si} + \lambda E_s \tag{18}$$

where A_i is the plan area of the ith leaf, $\hat{A}$ is unit area (for dimensional consistency) and the summation is over both sides of all leaves above unit area of ground. The surface temperatures and humidities are again unknown. Canopy models must allow estimation of these values throughout the canopy.

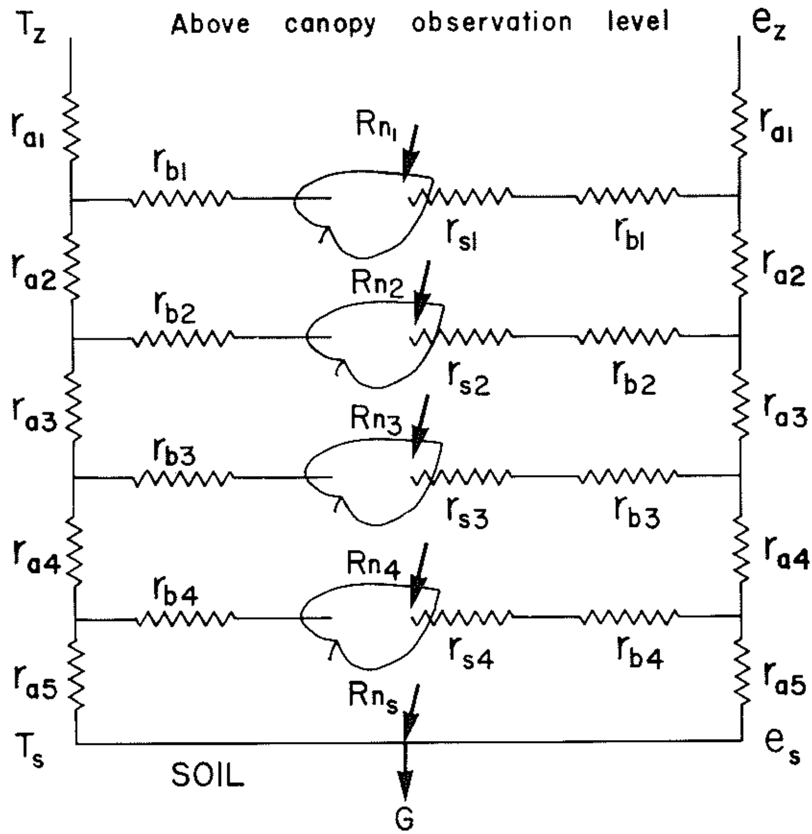

FIG.—1 Four-layered resistance analogue model of sensible heat flux pathways (left side of diagram) and latent heat flux pathways (right side of diagram) in a plant canopy. The two halves are linked by the energy balance requirement of the 'leaves'. Any number of layers can be incorporated. In the original version, Waggoner and Reifsnyder (1968) favoured nine.

The canopy models that have been developed are one–dimensional. Some of these are described in papers by Philip (1964), Cowan (1968), Waggoner and Reifsnyder (1968) and Lemon *et al.* (1971). Vertical profiles of leaf area, stomatal resistance, net radiation, wind speed and effective diffusivity are assumed known or are provided by various sub–models. Temperature and humidity conditions above the canopy are specified as well as the temperature and humidity or energy fluxes at the soil surface. With this information the models are solved for the profiles of temperature and humidity and profiles of the sensible and latent heat fluxes.

The model of Waggoner and his co-workers (Waggoner and Reifsnyder, 1968; Waggoner *et al.*, 1969) is representative. It is presented in the form of a resistance-analogue diagram in Fig. 1. By assuming that all leaves in a given layer have the same stomatal resistance and geometry, it is possible to describe evaporation from each layer by an equation analogous to (12) with the effective 'stomatal' resistance of the layer being found by dividing the average stomatal resistance of the leaves in that layer by the leaf surface area index of that layer. The model is solved as a set of linear equations (Furnival *et al.*, 1975). The Waggoner model simulates the profiles of temperature and humidity and the water vapour flux profile quite well. An example of measured and simulated profiles for a pine stand is given as Fig. 2.

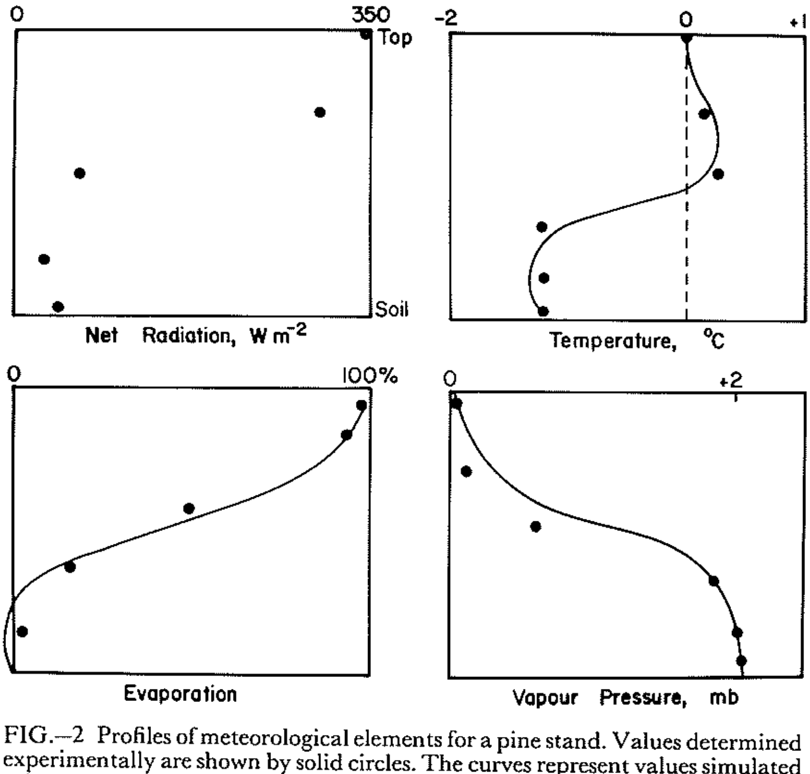

FIG.—2 Profiles of meteorological elements for a pine stand. Values determined experimentally are shown by solid circles. The curves represent values simulated using a nine-layered canopy model. Observations of temperature, humidity and evaporation are referred to conditions at the canopy top. (Redrawn from Waggoner *et al.* 1969).

The limitation of these models is the number and nature of the input data and the computing power needed to make the calculations. Rather than produce more detailed and more realistic models, practical applications call for simpler models that yet retain a reasonable ability to describe the energy exchanges of natural surfaces.

The simplest of these was developed by Penman and Schofield (1951) and Monteith (1965). If, in the multi–layer models, all of the intra-canopy aerodynamic resistances (r_{ai}) and boundary–layer resistances (r_{bi}) are assumed negligible with respect to the stomatal resistances (r_{si}), variations of temperature and humidity within the model canopy disappear. Equation (18) then reduces to

$$\lambda E = (\rho C_p / \hat{A}\gamma)\,(e_o^* - e_o) \sum_i A_i / r_{si} + \lambda E_s \tag{19}$$

A 'canopy' resistance (r_c) can then be defined by

$$r_c = 1 / \hat{A} \sum_i A_i / r_{si} \tag{20}$$

The canopy resistance is also known as the 'bulk stomatal' or 'surface' resistance.

Again in (19) the 'leaf surface' conditions are unknown. If soil evaporation be ignored, this canopy model is analogous to the single leaf model (12) and evaporation is related to conditions in the atmosphere above by an analogue of (17):

$$\lambda E = \frac{s(R_n - G) + \rho C_p (e_z^* - e_z)/\gamma r_a}{s + \gamma (r_c + r_a)/r_a} \tag{21}$$

where the aerodynamic resistance above the canopy, r_a, now appears in place of the leaf boundary-layer resistance r_b, and G is the ground heat flux density. Black *et al.* (1970) retained the soil evaporation term to obtain

$$\lambda E = \frac{s(R_n - G) + \rho C_p (e_z^* - e_z)/\gamma r_a - s + \gamma \lambda E_s}{s + \gamma (r_c + r_a)/r_a} + \lambda E_s \tag{22}$$

The aerodynamic resistance can be calculated from (11).

Considering its simplicity, the single-layer approximation performs well. Tests by Black *et al.* (1970), Brun *et al.* (1972), Szeicz *et al.* (1973) and Tan and Black (1976) encompass a range of canopies (snap beans, soybean and sorghum, sorghum, and Douglas fir forest, respectively) and give good agreement between model calculations and field measurements. A theoretical comparison with the multi-layer model by Shuttleworth (1977) shows close agreement between the two using values appropriate for a pine forest.

A point worthy of comment is that the single-layer canopy model, with soil evaporation included (22), has been tested and found reliable even for a row crop with two thirds of the area bare soil and a leaf area index of 1.25 (Black *et al.*, 1970). Evidently the model assumption of horizontal uniformity is not too critical. The theory seems adequate for sparse canopies, though the ability to measure or model soil evaporation is more important in these cases.

Naturally enough, the single–layer model breaks down in the case of a partially wetted canopy, for in that case the model assumption that the intra-canopy aerodynamic and boundary layer resistances are negligible relative to r_c is invalid because r_c is zero for wet leaves. An experimental illustration of this is given by Shuttleworth (1976).

The main conclusion to be drawn from these studies is that multi–layer models are available to mimic much of the observable meteorological behaviour of plant canopies. For larger scale or more practical studies the complexity of these models cannot easily be accommodated and the single–layer model provides a reasonable alternative.

Advective Enhancement

Canopy evaporation and the first phase of soil evaporation are strongly influenced by the vapour pressure deficit of the air close above. It is shown below that, for large uniform areas, the vapour pressure deficit near the ground is closely linked to the available energy and surface resistance. This linkage is broken when air with greatly differing charateristics is imported from another area. Evaporation rates may then be greatly enhanced or depressed. Two dimensional models must be used to examine these situations, which are characterised by horizontal variations in evaporation rates. These are called 'advection' models since they attempt to describe the horizontal changes in the heat and moisture content of the air and the horizontal variations in the convective fluxes.

To understand the effects of advective enhancement, it is useful first to consider a situation where advection has no effect. Consider a wind blowing across a very large, homogeneous, level region. To eliminate the effects of synoptic scale advection, suppose that the inversion layer, which typically caps the planetary boundary layer at about 1000 m, is impermeable. The surface fluxes of sensible heat and moisture must then be trapped within the planetary boundary layer and the temperature and humidity must continuously rise. If the surface is a plant canopy that can be described by (12) then a steady evaporation rate will eventually result when the vapour pressure deficit becomes constant:

$$\lambda E = [s/(s+\gamma)](R_n - G) \tag{23}$$

This is sometimes called the 'equilibrium' evaporation rate (McNaughton, 1976a).

The evaporation rate can then be described both by the equilibrium evaporation rate equation (23), and by the single–layer canopy model equation (12). Combining the two equations allows an 'equilibrium' vapour pressure deficit to be deduced:

$$e^* - e = [s\gamma/(s+\gamma)]\, r_c\, (R_n - G)/\rho C_p \tag{24}$$

'Equilibrium' evaporation does not necessarily occur from large, homogeneous regions, because the top of the boundary layer is not impermeable. Entrainment of drier air from aloft into the planetary boundary layer occurs (Tennekes, 1973), and should lead to evaporation rates somewhat larger than the equilibrium rate. However, if the latent heat flux were to become as large as the available energy the upward sensible heat flux would disappear, entrainment would largely cease, and evaporation rates would decline. An evaporation rate in the range

$$[s/(s+\gamma)](R_n - G) < \lambda E < R_n - G \tag{25}$$

is expected except where the evaporation rate is limited by water supply or marked local irregularities produce strong advective effects within the planetary boundary layer.

Some understanding of advection within the planetary boundary layer can be achieved by examining simplified models. McNaughton (1976b) has developed the model to examine the effects of a change in surface resistance or available energy on the local evaporation rate. Analysis shows that evaporation downwind of the boundary can be expressed by

$$\lambda E = [s/(s+\gamma)](R_n - G) + [s/(s+\gamma)]\,[r_c'(R_n' - G')/r_c - (R_n - G)]\;\Phi \tag{26}$$

where the primes signify upwind values and where Φ is an 'exchange function' that decreases from one to zero as distance inceases downwind from the boundary. The first term on the right is the equilibrium term which will be attained in the absence of all advective influences. The second term describes the advective enhancement (or depression) of the local evaporation rates.

If air blows from a dry region (large r_c') over a wetter area (small r_c) then advective enhancement will be large and positive – an 'oasis' situation. If the upwind area is wetter ($r_c' < r_c$) then (26) shows that the evaporation rate may be advectively depressed. Advective effects can also be induced by changes in available energy, but these will not often be large.

To estimate the mesoscale range of advective effects it is reasonable to ignore structure within the planetary boundary layer and treat the layer as vertically well mixed, with a constant wind speed throughout. Using these assumptions

$$\Phi = \exp\left\{-x\,(s+\gamma)\,/\,(H\gamma r_c u)\right\} \tag{27}$$

where H is the height of the planetary boundary layer and x is downwind distance. Using $H = 1000$ m and $u_* r_s/(s+\gamma) = 10$ (quite a low value), (27) gives a distance constant of 10 km. Advective effects will act over quite large distances, as has been observed experimentally (Burman *et al.*, 1975) and is illustrated in Fig. 3. Mixed layers developed over arid continental areas may be much deeper than 1 km and advective effects operate on an even larger scale. For smaller scale problems, account must be taken of boundary layer structure.

Some significant effect of advection on local evaporation rates is the rule rather than the exception. Provided these are not too large, the relationship (25) shows that good correlations between evaporation and available energy should obtain. This result is important and underlies the success of several evaporation estimation methods based on radiation correlations. Where advective enhancement is greater, or more accuracy is required, surface characteristics become more important and the theory gives little encouragement for the development of purely meteorological estimation methods. Models that include surface resistance values or some similar parameter must be used.

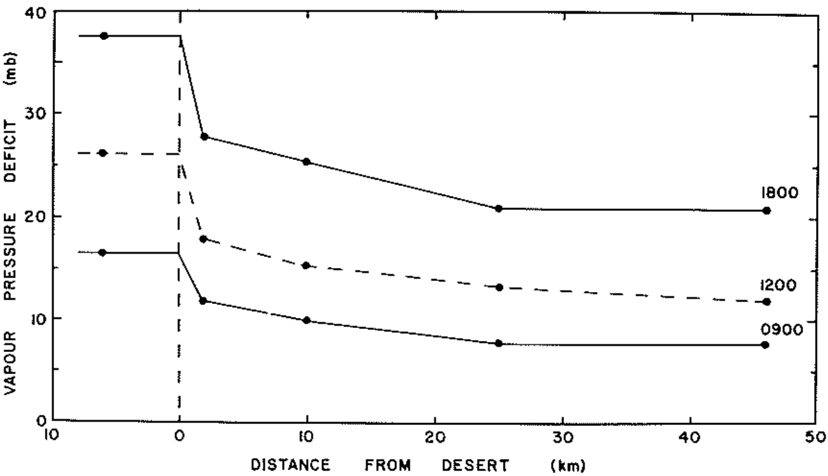

FIG.—3 Horizontal profiles of the vapour pressure deficit across from sagebrush desert into a large irrigated area in southern Idaho, USA. Initial adjustment is very rapid but changes still occur at distances greater than 20 km. (after Burman *et al.* 1975).

Evaporation from water–stressed vegetation

The one dimensional canopy models need no modification for the case of water–stressed vegetation, though higher values of stomatal resistance will generally obtain. However, when vegetation experiences water stress, stomatal resistances can be expected to vary with soil water potential and with the transpiration rate.

As a model input parameter, stomatal resistance presents some problems, as measurements must be made manually several times per day. On the other hand, soil water content, and hence soil water potential, can be estimated by soil water budgeting, with fewer field measurements. Quite naturally then, attempts have been made to develop relationships between stomatal resistance values and soil water status, or even to link transpiration rates directly to soil water content.

Denmead and Millar (1976) have made a detailed experimental study of plant resistances, water potential relationships and stomatal resistance in relation to transpiration for a wheat crop. They conclude that "there exists a maximum rate of transpiration that can be maintained by the plant – the rate equal to the flux of liquid water when leaf water potentials are at their critical values". If this is generally true, plant transpiration is similar to bare soil evaporation in that there are two separate phases; a meteorologically determined phase when stomata respond to radiation levels but not transpiration rate, and a water supply limited phase when stomata act to regulate a transpiration rate independent of meteorological conditions. The lesser of these two rates is the one that prevails at any given time. Transpiration during the second phase would depend on many plant factors (crop variety, stage of development, root development) which may vary widely. Other field results support this view of stomatal regulation of transpiration (van Bavel, 1967).

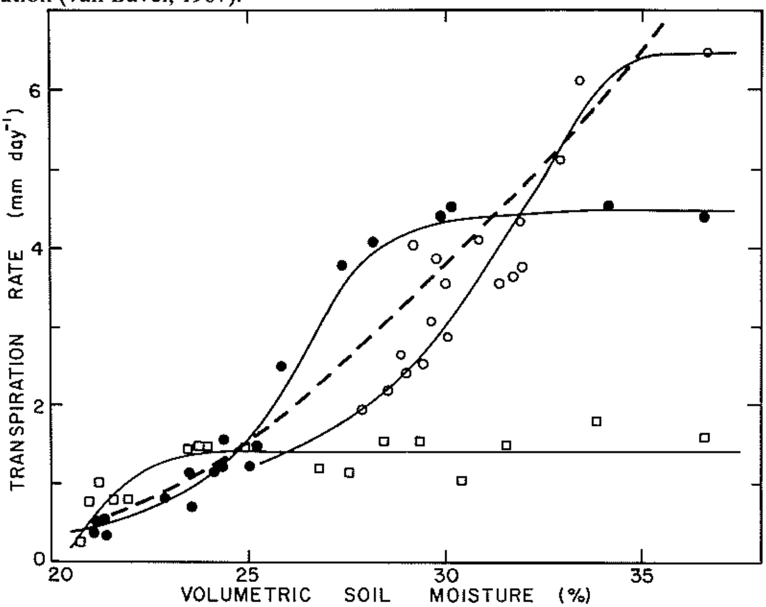

FIG.–4 Transpiration rates from pot-grown corn plants on three days (○ – clear, dry; ● partly cloudy, humid; □ heavily overcast, humid) against soil moisture content. The dashed line represents a water-supply-limited transpiration rate. At higher soil moisture contents the transpiration rate is independent of soil moisture (after Denmead and Shaw, 1962).

In the well known experiments of Denmead and Shaw (1962), stomatal response was not investigated, but transpiration rates from pot grown corn plants were related directly to soil moisture content. Some of their results, from three days with different weather conditions, are shown in Fig. 4. To a good approximation the downward sloping section of each curve can be approximated by the same dashed line, and evaporation is independent of meteorological conditions. (The position of the line should change as plants and roots develop). At higher soil water contents the curves approximate horizontal lines and transpiration rates are independent of soil moisture content.

The above two-phase model of evaporation from water–stressed vegetation has not been used in water budget models. Relationships based on experimental correlations have been more popularly accepted though they generally lack physical or physiological justification. These will be discussed in the section on evaporation estimation methods that follow.

ESTIMATION OF EVAPORATION

Direct measurement of evaporation over long time periods is difficult and methods of estimation are often used. Accurate estimates of evaporation from climatic data are needed in the analysis of irrigation requirements of crops, in hydrology and in crop production modelling. Ideally it would be feasible to estimate the evaporation rate for a given area using measurements of the climatic parameters in Monteith's equation (21) coupled with an estimate of the canopy resistance. However determination of r_c is not possible from climatic data and so evaporation estimation procedures have to rely on approximations or empirical correlations with climatic parameters to some degree.

Well-watered, Full-cover Crop Evaporation

For a well–watered crop with its canopy entirely covering the ground surface, growing in a region of small advective influence, the evaporation rate is dominantly controlled by the net radiant energy received as described by relationship (25). However soil heat flux averaged over 24 hours is small so reasonable estimates of evaporation can be made from net radiation data alone.

Net and solar radiation are closely correlated, therefore evaporation from well–watered full–cover ps is related to the more easily measured solar radiation (Jensen and Haise, 1963; and Jensen *et al.*, 1970). Makkink (1957) found in Holland that:

$$\lambda E = 0.61\,[s/(s+\gamma)]\,R_S - 3.4 \tag{28}$$

Priestley and Taylor (1972) found when advection is small evaporation is strongly correlated with the equilibrium evaporation rate as follows:

$$\lambda E = \alpha\,[s/(s+\gamma)]\,R_n \tag{29}$$

where $\alpha = 1.26$ ($\pm 5\%$). *The limits of* α can be deduced from (25),

$$1 < \alpha < (s+\gamma)/s \tag{30}$$

At 20° C, $(s + \gamma)/s = 1.46$. Several studies over various well–watered full cover crops have found α to vary over the range $(1, (s + \gamma)/s)$ with a mean of about 1.2 (Tanner and Jury, 1976).

We have calculated α from daily Bowen ratio-energy balance evaporation measurements made at Palmerston North during late spring–summer over pasture, paspalum and lucerne (Kerr *et al.*, 1973; Kerr, J. P. and J. S. Talbot, unpublished data) and winter–early spring over oats (Clothier, 1977). Using the warm season data the regression of measured evaporation on the equilibrium

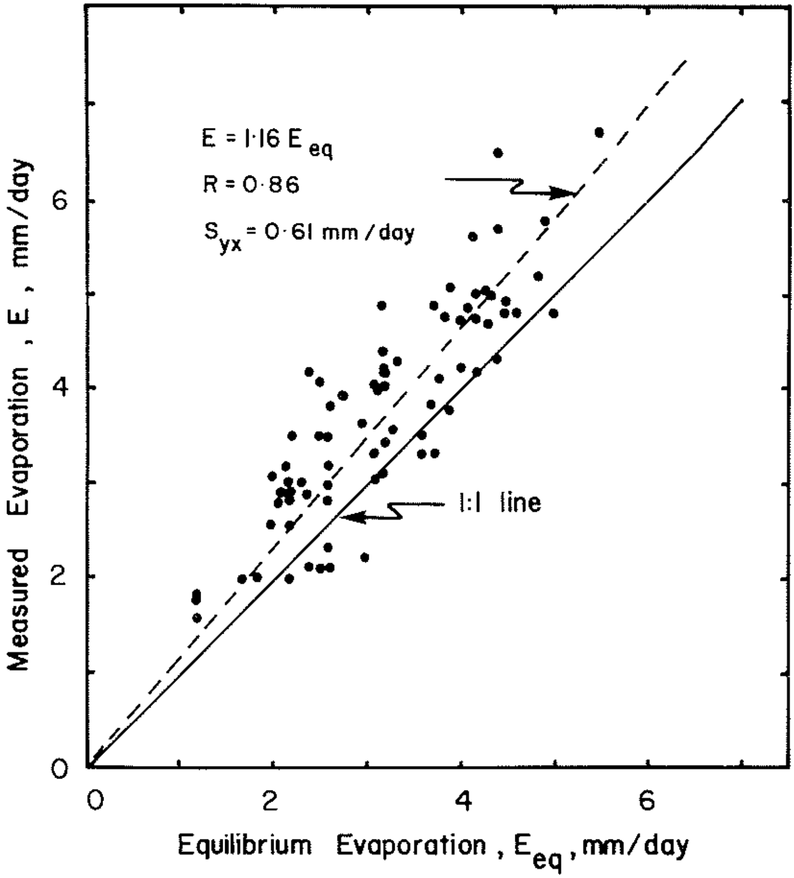

FIG.—5 Comparison of measured daily evaporation (E) by the Bowen-ratio energy balance technique, and the equilibrium rate (E_{eq}) for lucerne, paspalum and pasture near Palmerston North. The standard error of the estimate (Syx) relates to the regression line constrained through the origin.

rate produced $\alpha = 1.16$ (Figure 6). For the cool season $\alpha = 1.22$. The combined data yield $\alpha = 1.21$. These data relate to daylight measurements only. For 24 hour measurements α is larger, especially during winter when R_n is low. An α value of 1.21 is similar to many overseas observations.

The Priestley and Taylor method produced daily and weekly estimates of evaporation with an error of ± 15% and ± 5% respectively, using the Palmerston North data. The method provides a useful means of estimating evaporation from well–watered, full–cover surfaces, in the absence of significant advection.

These radiation correlation methods of predicting evaporation cannot be universally applied because advective effects will vary between crops and locations. There will also be a residual variability not correlated with radiation and local calibration of empirical evaporation estimation methods is necessary (Pruitt and Doorenbos, 1977).

Some evaporation estimation formulae have attempted to account for the advective fraction of evaporation. Penman (1948) developed an equation for estimating evaporation from wet surfaces. His equation can be obtained by setting $r_c = 0$ in (21) or (22) so that:

$$\lambda E = [s/(s+\gamma)](R_n - G) + \rho C_p (e_z^* - e_z)/(s+\gamma)r_a \qquad (31)$$

Using data on evaporation from a sunken pan, Penman found that:

$$r_a = 250/(1 + 0{\cdot}54\,u) \qquad (32)$$

where u is windspeed in m s^{-1} at 2 m. By comparison of (31) with (26), the second term of Penman's equation can be interpreted as an advective enhancement term for evaporation from wet surfaces.

Later work by Businger (1956), Thom (1972) and Thom and Oliver (1977) showed that r_a should be calculated from wind profile theory, and that (32) holds

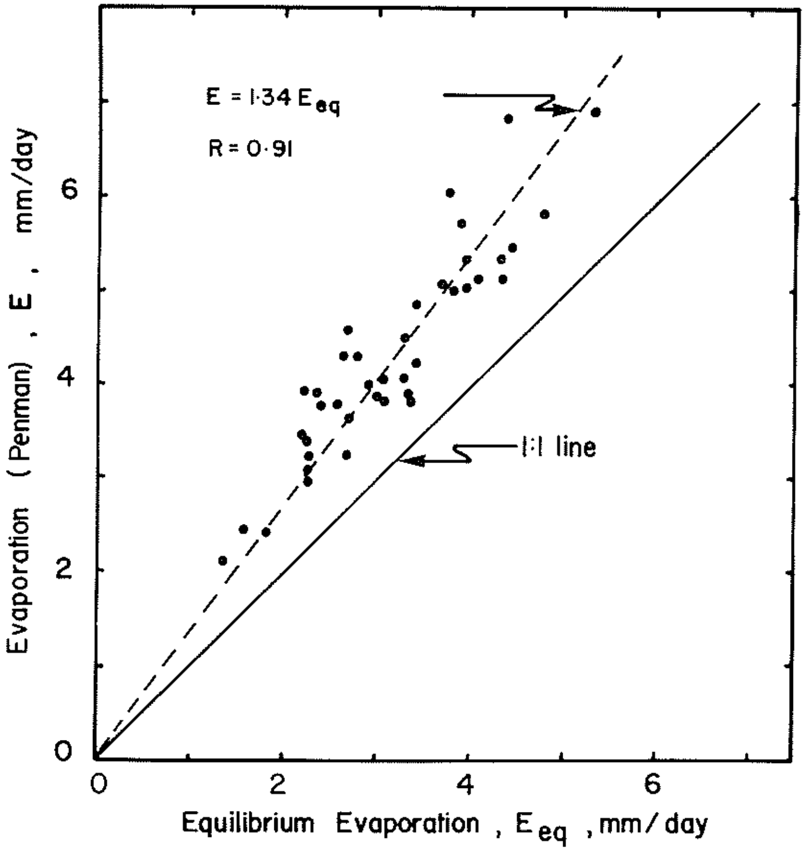

FIG.—6 Comparison of evaporation estimates based on Penman's equation (E) with the equilibrium evaporation rate (E_{eq}) for 40 days in January to March 1972. Data for lucerne, paspalum and pasture at Palmerston North.

only for surfaces with very small roughness lengths. Thom and Oliver argue that when Penman's equation is used to estimate evaporation from vegetation the overestimate of r_a is largely compensated by the neglect of r_c in (21). For European conditions Penman (1956) noted that the ratio of the 'equilibrium' term to the advective term was usually in the range of four or five to one. That the advective term is often small, and its empiricism contains compensating errors means Penman's equation often provides reasonable estimates of evaporation.

Penman's estimates of evaporation are compared with 'equilibrium' evaporation estimates for data collected in summer 1971/72 at Palmerston North in Fig. 6. Penman's estimate of advective enhancement is 25%. However measurements of evaporation (Fig. 6) indicate that on average advective enhancement increases evaporation by only 16%. Penman's equation, in common with all other empirical methods, can be improved by local calibration.

In regions of small advective influence the Priestley and Taylor approach appears the most useful evaporation estimation method, as the extra data and

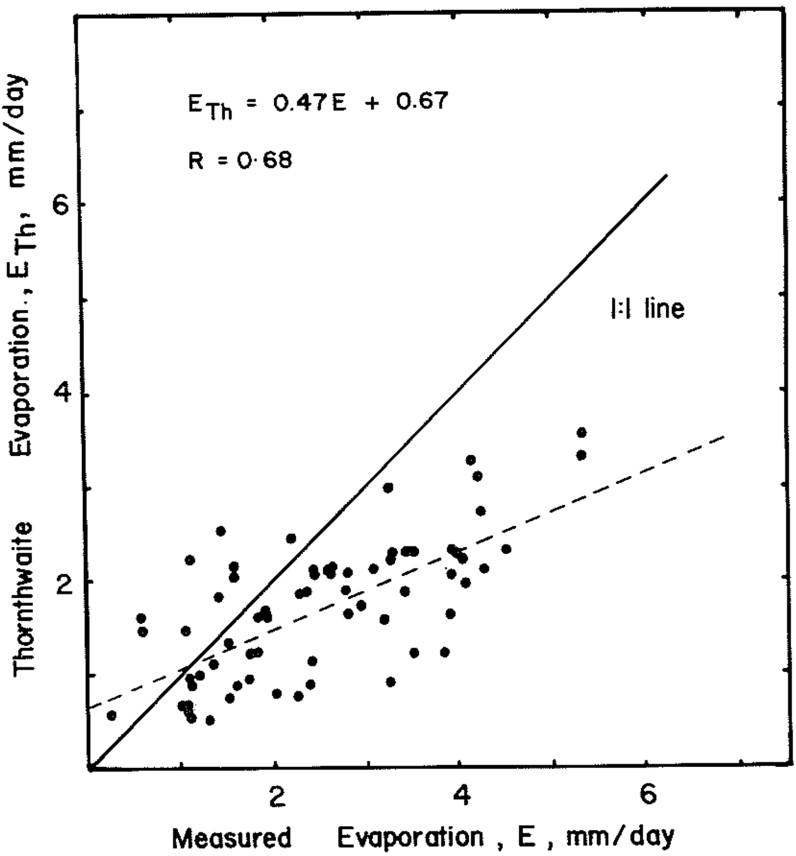

FIG.—7 Comparison of evaporation estimates based on Thornthwaite's procedure (E_{Th}) with evaporation (E) measured over autumn–sown oats, by the Bowe, nratio energy balance technique for 69 days during 1974.

calculation complexity needed for Penman's equation is not rewarded by a better estimate of evaporation.

Pans are often used to estimate evaporation. Penman's equation (31) applies to pans if R_n is interpreted as the pan net radiation, G is interpreted as heat loss through the pan walls and r_a is interpreted as the heat and mass transfer resistance from the pan. Regression relations between pan evaporation and land evaporation will depend markedly on pan design and exposure and climatic conditions. Pruitt and Doorenbos (1977) using a lysimeter reported that the ratio pasture to Class A pan evaporation varied from 0.43 to 1.05 depending on both the nature of, and distance to the immediate upwind surface and also on the prevailing climatic conditions. They conclude that "as with other methods hardly enough stress can be placed upon the importance of local calibration". Evaporation pan data are readily available. In New Zealand, Coulter (1973) reported that only 9 meteorological stations recorded solar radiation whereas 38 possessed evaporation pans.

A much less satisfactory method of estimating evaporation is that of Thornthwaite (1948), which is based on an empirical correlation between temperature and evaporation. Air temperatures are usually correlated with net radiation and this is the physical basis of its success, but these regressions are far from consistent, particularly in the short term, and often explain only a small part of the variance. An example of poor performance of the Thornthwaite method is given in Fig. 7. The method has been extensively used by workers in New Zealand (Rickard and Fitzgerald, 1970), largely due to the lack of information other than temperature data (Lamb, 1974). Its use should be confined to these cases and its very great limitations recognised.

Estimation with Strong Advective Enhancement

If advective enhancement is large then it becomes more important to estimate this enhancement accurately and radiation correlation based methods are less valuable. Calculation directly from the advection model (26) is not feasible. Though already complex, this model is still too crude for practical use. The Penman-Monteith model (21) offers another starting point that would be useful if r_c values were known to sufficient accuracy. This method does not appear to have been used successfully.

In practice, the same methods have been used to estimate evaporation when advective enhancement is large as at other times. Penman's equation in its traditional form has often been used to estimate evaporation in advective conditions. The accuracy of the advective term in such regimes has usually been found to be unsatisfactory (Thompson and Boyce, 1967; Rosenberg, 1969). This has lead to further empirical modification of the advective term, such as Greenland's (1971) attempt to account for the Canterbury nor'wester. In view of the popular opinion to the contrary, it must be reiterated that the advective term in Penman's equation is quite empirical and must be calibrated locally for good results. Tanner and Jury (1976) have proposed an advective correction to the Priestley and Taylor formula, but the limitations previously mentioned apply and we conclude that presently there is no entirely satisfactory method of estimating advective enhancement.

In extreme cases strong advective enhancement and low radiation levels may coincide. In Canterbury this can occur when a strong dry nor'wester blows but the land is shaded by the nor'west arch. None of the present estimation methods will give reliable estimates in these circumstances.

Evaporation During Crop Water Stress

It is often necessary to estimate evaporation from water–stressed vegetation. Evaporation rates fall below those expected for similar unstressed vegetation and may become independent of meteorological conditions, as discussed earlier.

Application of a two-phase model of evaporation is not yet possible because of the difficulty in calculating the rate of evaporation during the plant limiting phase. Usually reduced rates of evaporation are computed by multiplying a supposed value for the well–watered evaporation rate (E_{max}) by a coefficient (h) that attempts to account empirically for soil water stress. This can be written as:

$$E = h\, E_{max} \tag{33}$$

E_{max} is usually defined by one or other of the evaporation estimation procedures. It is a 'fictitious' evaporation rate since in this context it never exists.

The most widespread empirical approach is to consider the ratio E/E_{max} as a function of the available soil water storage (θ^*) in the root zone, i.e. where:

$$\theta^* = \frac{\theta - \theta_w}{\theta_f - \theta_w} \tag{34}$$

and θ is the water content, θ_f the 'field capacity' water content, and θ_w the lower limit plant available water content. In this way meteorological and soil factors are supposed to act together to determine the evaporation rate. This is conceptually quite different from the two-phase model proposed ealier where only one of these is effective at anytime.

Two approaches have been used to describe the decline in the relative evaporation rate with θ^*: those using one of many non–linear functions over $0 \leq \theta^* \leq 1$ and those which define a critical storage value θ_c^* such that $h = 1$ for $\theta^* \geq \theta_c^*$ and declining in some way to zero as θ^* approaches zero. The latter approach has been most widely applied, generally with $0.25 \leq \theta_c^* \leq 0.35$ and a linear decline below θ_c^*. Obtaining water–stressed, full–cover evaporation from such an approach with $\theta_c^* \approx 0.3$ appears to be a useable empiricism, although it lacks a satisfying fundamental basis (Tanner, 1967).

Equation (33) need not lead to large errors when used in water budget models. Early overestimates of E would lead to a more rapid computed depletion of available soil water and hence subsequent estimates would compensate by being too low. Accumulated evaporation is limited by the available soil water.

Water budget calculations are susceptible to poor estimates of the 'field capacity' θ_f and the lower limit of plant available water θ_w. Veihmeyer and Hendricksen (1931) concluded that laboratory estimates of θ_f, at about $-\frac{1}{3}$ bar potential, "indicate the field capacities of deep, well–drained soils with no decided changes in texture or structure at least for the fine-textured soils". However, the concept of 'field capacity' has been flagrantly misapplied, and the only means of determining θ_f in most field soils is by field sampling following wetting–up of the profile to determine when a quasi–static equilibrium has been achieved, after say 2–3 days. For Manawatu fine sandy loam a soil underlain by coarse gravels Clothier *et al.* (1977) found the $-\frac{1}{3}$ bar water content underestimated the *in situ* value of storage by 95 mm. The lower limit of plant available water θ_w is commonly estimated using the -15 bar water content but the steepness of water retentivity curves in this region means that the exact choice is not critical. Error in assessing rooting depth can also introduce significant errors with water budget calculations based on (34).

Evaporation From Sparse Canopies

Soil evaporation comprises a significant fraction of the total evaporation from crops during their establishment stages, and from other sparse canopies. Though (22) has been used in research studies, the difficulties in knowing values for r_c again arise. A practicable approach is to use separate estimates for soil evaporation and transpiration but to use empirical methods to estimate these components.

Ritchie (1972) has developed an elaborate computer water budget model that includes separate calculations of soil evaporation and transpiration. Soil evaporation is calculated from a two-phase model using:

$$\lambda E_S = [s/(s+\gamma)]\, R_n \exp(-0.4\, LAI) \tag{35}$$

to estimate evaporation in the meteorologically limited phase. In (35) the exponential factor reduces above–canopy net radiation to a soil–surface value. After a fixed amount of water has been removed from the soil a square–root–of–time relationship is used for the second phase drying.

To estimate transpiration the relationship:

$$E_T = g\, E_{max} \tag{36}$$

has often been used. Ritchie and Burnett (1971) set E_{max} equal to the net radiation intercepted by the canopy and found experimentally:

$$g = -0{\cdot}21 + {\cdot}70\, LAI \text{ for } 0{\cdot}1 \leq LAI < 2.7 \tag{37}$$

Ritchie (1972) used this relationship in his model. Separate calculations of soil evaporation and transpiration were not made when the leaf area index exceeded 2.7. Ritchie ignored the interdependence of soil and canopy evaporation that is predicted by physical models such as (22).

Tanner and Jury (1976) used Ritchie's methods but found that they overestimated evaporation from potatoes by 10%. Small modifications to the coefficients (local calibration) gave a better fit between the water budget calculations and field measurements. They concluded that the "simple procedures appear useful for separately estimating evaporation and transpiration provided transpiration is not limited by soil water shortage".

Water stress may occur in sparse–canopied vegetation. For desert plants this is quite common but may also occur with crop plants when drought follows sowing. For this situation some modellers use a relationship in the form

$$E = g\, h\, E_{max} \tag{38}$$

Since evaporation rates should be very low at such times, one can only hope that relative errors resulting from this procedure will have little importance.

GENERAL COMMENTS

Very few direct measurements of evaporation have been made in New Zealand. No reliable lysimeter measurements of evaporation have been published. Energy balance/Bowen ratio measurements of evaporation have been made at Palmerston North and Lincoln where several hundred days' data have been collected, over a variety of crops and pasture but much of this is unpublished. Evaporation can be computed as a residual term in soil water budget calculations and the results compared with results from evaporation formulae (Fitzgerald and Rickard, 1960). However the methods depend on accurate estimates of drainage and soil water storage, and errors are hard to avoid. Daily values cannot be calculated by this method (Rouse and Wilson, 1972).

With these exceptions local calibrations of evaporation estimation methods are not available in New Zealand and overseas information must be relied on. New Zealand has a sub–humid climate and in similar climates overseas Penman's equation has been extensively used and found reasonably reliable. From available data including that presented above, it appears that the simpler Priestley and Taylor equation (with $\alpha = 1.2$, say) gives results of comparable accuracy. Use of these methods depends on the availability of radiation data. A denser network of radiation instruments will be needed before they can be fully utilized in New Zealand. Until this occurs, data from the many evaporation pans will continue to be used. No direct determinations of pan coefficients have been published in New Zealand. Where advective enhancement is not small, lack of local calibration of methods used in New Zealand means that evaporation estimates have quite large uncertainties. This is unfortunate, particularly in view of the development of irrigation schemes in Canterbury, the inland South Island basins and elsewhere where advection may be significant.

Looking ahead, an increased use of evaporation estimates for on–farm scheduling of irrigation can be expected. Water supplies for irrigation are often limited, and costs of application are a significant factor in the development of sprinkler irrigation. The most successful scheduling methods are based on determining crop evaporation, measuring the soil water–holding capacity within the crop root zone plus rainfall, and from this information computing the crop water budget (Jensen *et al.* 1971). When soil water falls below a nominated level irrigation water is applied so that the crop growth is not restricted by water stress.

Evaporation estimates are usually based on point measurements at present. Estimates over extended areas of crops and pastures, catchments, or similar large scale vegetation units would be useful. The possibility of using satellites for estimating evaporation from these extended areas has considerable prospect and is being examined currently. Satellites can be used to measure the leaf area development of a crop (Kanemasu, 1974) and aircraft used for remote sensing of canopy surface temperatures. Satellite information on the crop leaf development can be combined with climatological observations to compute the evaporation from a particular region. The method is limited at present because of the delays in collecting data and relaying to ground observers.

It should also be feasible to calculate net radiation values from satellite data and to use these to make estimates of evaporation using the Priestley and Taylor formula. Future developments in this area will lead to improved ability to calculate evaporation from large and remote areas of various topographies where few data are currently available.

A number of topics worthy of further research can be identified. The lack of reliable values for canopy resistance places limits on the potential accuracy of any evaporation estimation method. Research towards providing suitable values is potentially very useful, though worthwhile results will probably be difficult to obtain. Correlations between r_c and meteorological parameters have been obtained (e.g., Monteith, 1965) but the relationships are not physiologically based and rely on conditions at the sites where the base data are collected. To provide the extra information needed r_c should be related to the plant canopy characteristics such as leaf area index and stomatal resistance. Problems of data collection and model fidelity will have to be examined. No satisfactory solution to these problems is yet in sight.

A second research problem relates to a topic that has not been discussed above. Both natural and man made alterations to the landscape may have impacts on evaporation rates. Farmers use windbreaks, sometimes with the intention of reducing evaporation rates and promoting crop yield. Afforestation of previously open catchments or other major changes in vegetation can alter evaporation rates significantly with consequent changes in stream flow. These situations cannot be usefully modelled using one dimensional models and must be treated using more complex advection models. To date these have been developed only for very simple experimental situations.

Broken topography is one of the outstanding characteristics of New Zealand's agricultural and forestry lands. Approximately 80% of the land area comprises steep and hilly regions with slopes greater than about 12° (Leamy, 1974). Evaporation estimates are often needed, either for individual catchments or for a family of catchments covering an extended area. Within individual catchments net radiation varies considerably over the range of slopes and aspects, and the eddy transport processes which operate cannot be adequately described using simple one dimensional models developed for flat regions. These effects on evaporation are unknown.

Even were these scientific questions answered, a substantial educational problem would remain. Reliable information on evaporation rates and estimation procedures needs to reach the hands of users such as design engineers, farm advisory officers, consultants and the like. Judging by overseas experience, this problem is almost as difficult to solve as the purely scientific problems.

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SYMBOLS AND UNITS USED IN THE TEXT

A	Leaf area	m^2
$\bar{A}$	Unit area	m^2
C	Sensible heat flux density	$W\ m^{-2}$
D	Hydraulic diffusivity	$m^2\ s^{-1}$
E	Evaporation rate	$kg\ m^{-2}\ s^{-1}$
E_{max}	Supposed maximum evaporation rate	$kg\ m^{-2}\ s^{-1}$
E_S	Evaporation rate from the soil surface	$kg\ m^{-2}\ s^{-1}$
E_T	Transpiration component of total evaporation	$kg\ m^{-2}\ s^{-1}$
G	Ground heat flux density	$W\ m^{-2}$
H	Height of the planetary boundary layer	m
K	"Effective" or "Eddy" diffusivity	$m^2\ s^{-1}$
LAI	Leaf area index	—
R_n	Net radiation heat flux density	$W\ m^{-2}$
R_S	Short-wave radiation heat flux density	$W\ m^{-2}$
T	Temperature	K
C_p	Specific heat of air at constant pressure	$J\ kg^{-1}\ K^{-1}$
e	Vapour pressure	mb
e^*	Saturation vapour pressure	mb
g	Reduction factor in (36)	—
h	Reduction factor in (33)	—
m	Proportionality constant in (2)	$kg\ m^{-2}\ s^{-1/2}$
q	Absolute humidity	$kg\ m^{-3}$
r	Transport resistance	$s\ m^{-1}$
s	Rate of change of saturation vapour pressure with temperature	$mb\ K^{-1}$
t	Time	s
u	Wind Speed	ms^{-1}
u_*	Friction velocity defined by (9)	ms^{-1}
x	Horizontal distance	m
z	Height	m
Φ	Exchange function	—
α	Proportionality constant in (29)	—
γ	Psychrometric constant	$mb\ K^{-1}$
θ	Soil water content	$m^3\ m^{-3}$
θ_f	Soil water content at "field capacity"	$m^3\ m^{-3}$
θ_w	Soil water content at permanent wilting point	$m^3\ m^{-3}$
θ^*	Available soil water defined by (34)	—
κ_E	Molecular diffusivity of water vapour in air	$m^2\ s^{-1}$
λ	Latent heat of vaporisation	$j\ kg^{-1}$
ρ	Density of air	$kg\ m^{-3}$
τ	Momentum flux density or shearing stress	$kg\ m^{-1}\ s^{-2}$

Additional subscripts have the following significance

a	aerodynamic
b	boundary-layer
c	canopy
i	for the ith leaf or layer
s	stomatal
z	at height z above the surface
o	at the surface
C	for sensible heat transport
E	for vapour transport
τ	for momentum transport

Primes denote upwind values

SLOPE STABILITY STUDIES IN NEW ZEALAND

M. J. Selby

ABSTRACT

Over 130 papers on slope stability studies have been published in New Zealand. The great majority of these are concerned with descriptive accounts of the relationship between deforestation, or vegetation, and erosion by landsliding. Nearly all parts of the upland areas of the country are affected by instabilities at scales ranging from huge tectonic movements of ranges in mudstone rocks to common translational slides in regolith materials. The use of field shear boxes is starting to throw light on the effect of vegetation on soil strength, and the use of dated volcanic ash beds as markers is permitting a chronology of erosional episodes to be constructed. Determination of the magnitude and frequency of storm events has aided calculation of denudation rates. The use of a standard terminology, classification and system of reporting landslides is advocated. Future studies are likely to involve the use of geotechnical methods, a concentration on instability in urban areas and closer study of structural and lithological influences on instability.

INTRODUCTION

The phrase 'slope stability' is usually used to refer to mass wasting and consequently excludes that group of slope processes resulting directly from the action of sheet wash and channel erosion on slopes. Such a distinction is inevitably arbitrary because many mass wasting scars and deposits become sites for the initiation of rills and gullies, but discussion in this account will be confined to the processes and forms of mass wasting of slope materials which range from slow creep to landslides and mass flows.

Landslides are some of the most obvious features of the upland landscape of New Zealand. Many factors contribute to their abundance: high available relief; active tectonism; main axial mountain ranges formed largely of highly folded, faulted and shattered greywacke; large areas of Cenozoic rocks with a high clay content; deeply weathered regoliths in the North Island; weakly consolidated eolian cover beds of volcanic and loessic origin; steep slopes; vigorous undercutting of slopes in many areas of fine fluvial dissection; and vegetation cover greatly modified by human activity in the last thousand, and especially the last hundred, years.

Landsliding has attracted considerable attention and prompted a substantial literature. Approximately 130 scientific papers and reports have been published (all known reports to 1975 are listed with brief abstracts in Selby, 1976a) and many pamphlets, newspaper accounts and brief descriptions have been produced. Of the 130 papers: 11 discuss the distribution of mass wasting; 13 are concerned with its magnitude and frequency of occurrence; 16 with its quantitative and economic significance; 28 with lithological and pedological conditions; 32 with trigger mechanisms; 10 with methods of investigation; 20 with the morphology of mass wasting; 13 with the rates of mass wasting processes; 60 with types of processes; 6 with methods of classification; 6 with remedial action and over 120 with the relationship between mass wasting and vegetation cover – particularly with the results of changes in vegetation. Many papers are, of course, concerned with more than one aspect of study but the importance attached to vegetation emphasises the concern that has developed over the effect of human interference with the original forest, which formerly covered most of New Zealand below the tree line.

The only comprehensive review of mass wasting processes, their morphology and distribution, is that of Cumberland (1947). This pioneer work brought the extent of soil erosion problems to the attention of a wide audience and, together with the reports of Gibbs and Raeside (1945) and Grange and Gibbs (1947), prepared legislators, farmers and the general public for the development of a national attack on the problems of maintaining land in production while reducing erosion and protecting soil from potential instability. The history of soil conservation measures in New Zealand has been reviewed by McCaskill (1973).

MODE AND LOCATION OF MASS WASTING

One of the main concerns of Cumberland's book was to describe the variety and extent of mass wasting and erosion processes and their regional significance. His descriptions have not been superseded and comment here is confined to work or events which have occurred since his book appeared.

The eastern fold belt of the North Island is a zone of folded and faulted sedimentary rocks, of late Cretaceous to Pliocene age, forming the hill country in land of the east coast. This belt has many examples on its slopes of instabilities ranging in scale from large tectonic features in which whole ranges have moved and whole valleys been formed by massive landsliding, to small scale landslides affecting a few tens of square metres.

In the coastal zone of southern Hawke's Bay the whole landscape over a width of 3 to 5 kilometres, and a north to south distance of 20 km is involved in a complex and fragmented regional slump (J. R. Pettinga, University of Auckland, pers. comm.). Large disturbances in Hawke's Bay and Poverty Bay involve many lines of weakness, some of which have bentonitic intrusions, and relatively weak mudstones and siltstones have been extruded or flowed from under caprocks of sandstones and limestones. Such features are common in both the Raukumara Peninsula (M. Gage, Canterbury Museum, pers. comm.) and southern Hawke's Bay. Many intermediate scale features occupying a few hectares owe their origin partly to tectonic activity, and the formation of zones of shattered rock which can be readily penetrated by water, and partly to the montmorillonite and illite clays within the sediments which swell upon wetting (Claridge, 1960).

The main axial ranges of the North Island, stretching from Wellington to Whakatane, are formed of highly folded and shattered greywacke, often overlain by weakly consolidated volcanic ashes. These ranges are raised to over 1000 m in many areas and they have steep slopes and deeply incised valleys. Landsliding is mostly of the translational slide type and penetrates the bedrock as well as involving the ash mantle. In both the northern and southern parts of the Ruahine Range erosion is very severe (Grant, 1965; Stephens, 1975; Mosley, 1977), but much of the concern about this has been directed at the aggradation of stream channels in the headwaters, and the effects this may have downstream, rather than at the landsliding and gully erosion itself.

At the southern end of the greywacke ranges storms in the 1970s have caused considerable landsliding and flooding in the settled area of Wellington and the Hutt Valley and directed much attention to slope stability problems in urban areas (Eyles *et al.*, 1974; Taylor *et al.*, 1977).

The Auckland metropolitan region has also suffered some instability problems where a complex geology of volcanic and sedimentary rocks of the Waitemata Group has formed a mosaic of stable and potentially unstable slopes. The Waitemata Group rocks consist mainly of alternating layers of soft mudstone and harder sandstone and 'softening' of the mudstone, especially along fissures, has given rise to many failures, often at very low angles of slope (Taylor *et al.*, 1977).

The South Auckland greywacke ranges of Hunua and Hapuakohe have received, and are still receiving, considerable attention. They were affected by a series of summer rainstorms which produced many shallow landslides in the 1960s and 1970s (Selby 1967a, b; 1976; Pain, 1968, 1969, 1971). These ranges have a regolith of deeply weathered greywacke and a mantle of weathered volcanic ashes. It is the weathered colluvial materials which are particularly prone to movement.

Taranaki and Wanganui provinces are largely underlain by interbedded grey mudstones, siltstones and sandstones, known collectively as 'papa'. They constitute 'soft' rocks in the engineering sense ('soft' rock being a natural aggregate of minerals which are bonded together such that, unlike a soil, they cannot be separated by gentle agitation in water). The height and steepness of the slopes in 'papa' country indicate that the rock has considerable inherent strength and deep–seated rotational slumps are not known to occur even where river undercutting has produced cliffs up to 150 m high (Hawley *et al.* 1975). Various types of translational slide – especially creeping earthflows and shallow debris avalanches are, however, common. The creeping earthflows are of particular economic significance because they increase the costs of maintaining roads and railways. The large slide–flow at Utiku on the North Island Main Trunk Railway began moving in 1964 (Belz, 1967) and caused the railway track to drop at the rate of 8 cm a day. In the last quarter of 1965 it caused horizontal movement of about 1 cm a day. Drainage of the flow has reduced rates of movement but not stopped them (Ker, 1970).

In the Wairarapa, Blenheim area and the Port Hills south of Christchurch, loess mantled slopes have suffered considerable mass wasting. In the Wairarapa the very wet winter of 1977 caused many debris flows and debris avalanches. The numerous shallow landslides in the hills east of Masterton resulted from saturation of the regolith. The failure planes occurred at the base of the solum and above the underlying siltstones. The yellow grey earth soils, formed in late Pleistocene loess, reached their liquid limits in many places, and flowed to leave large scars on the hills and low angle fans of saturated colluvium on the valley floors.

In the South Island loess areas one of the characteristic forms of erosion is tunnel–gullying (Gibbs, 1945; Hosking, 1962; Hughes, 1972; Laffan *et al.*, 1977). Such erosion occurs primarily on north–facing slopes in soils which become cracked during a dry summer. With percolation of water down the shrinkage cracks during a succeeding wet season the soil lining the crack disperses and cavities form. Enlargement and linking of the cavities allows tunnels to develop either along the upper surface of the fragipan of yellow grey earths or below the fragipan. Tunnel roofs may collapse and so give rise to gullies, and the channelling of drainage may produce local soil saturation and then landsliding. Landslides on the loess also develop independently of tunnel gullies and are frequently associated with soils which crack in summer and then fail in wet seasons when water can enter the soil through the cracks. The spread of settlement on the Port Hills in recent years, and the consequent threat of instability to property, has directed attention to this form of erosion.

The Kaikoura Coast of the South Island is renowned for the extremely high rainfall which can occasionally fall in a short period. One storm, cyclone Alison, in March 1975 caused severe landsliding of complex slide–avalanche–flow type and consequent aggradation in the steep coastal catchments (Bell, 1976a). Similar results of infrequent high intensity storms have been described from further south in the Canterbury Alps: a severe storm in the Godley Valley on 26 December 1957, when 480 mm of rain fell in one day, produced massive landslides and valley floor aggradation (Scott, 1963). Buried soils in valley floor deposits and the stability of vegetation on fan surfaces indicate that events of similar magnitude occur each century.

In the Otago schist zone of the South Island mountains many slopes are prone to failure where valley walls have been oversteepened by glacial erosion and stream undercutting. In areas like the Kawarau Valley (Bell, 1976b) and Matukituki Valley failure surfaces may be up to 100 m below the ground surface. They occur commonly in planes parallel to the planes of schistosity and are particularly common where seams of clay occur within the schist – possibly also along lines of micaceous foliae and along zones of weak talc schist. Shallow failures in the periglacial deposits mantling many slopes are also common. Many valley sides, therefore, have numerous mass movement features of a great range of ages, a variety of sizes and many modes of failure.

The Otago Peninsula exhibits a variety of slope failures. Deep rotational slumps in Cenozoic sedimentary rocks were described by Benson (1940, 1946) who discussed their importance for railway alignment and the location of buildings. Earthflows and compound slide–flow features in loess and periglacial debris have been discussed by Crozier (1968, 1969) and by Leslie (1974) who related soil type, landslide hazard and land use planning requirements.

On the West Coast of the South Island the Fiordland area has not been examined for its slope stability characteristics. The massive igneous and metamorphic rocks may well be resistant to slope failures although some unloading phenomena, following deglaciation, might be expected to occur. The shallow failures, of debris avalanche form, which result from loading of very thin regoliths by a forest cover, and the consequent vegetation cycle of bare rock to lithophytes to scrub and then forest cover have been described by Wright *et al.*, (1952) and by Mark *et al.*, (1964).

Farther north the very steep slopes formed in less resistant rocks, greywacke and schist, which are still undergoing tectonic uplift in a super–humid climate, are prone to shallow landsliding of the debris slide and debris avalanche type. This region is also affected by periodic earthquakes, like that of 1968 (Adams *et al.*, 1968) in the Inangahua area when a shallow earthquake of magnitude 7 caused many slope failures within a radius of 16 km of the epicentre. High lateral and upward accelerations caused many block glides as well as shallow debris avalanches in shattered rock and saturated soils (Johnson, 1974).

TOPICS OF STUDY

Urban Development

The spread of settlement in the major urban areas on to areas of potential slope instability, and the effects of a number of severe storms in the Wellington and Hutt Valley suburbs as well as the Port Hills, has provoked considerable interest during the 1970s. The requirements upon local government authorities to ensure that housing sites are properly evaluated for stability, before building licences are issued, has also drawn a number of consulting engineers and engineering geologists into this field of study (Taylor *et al.*, 1977).

A contentious issue arising from storm damage on 20 December, 1976, in the Wellington area is the question of how far cuts and fills, made to produce platforms for houses, increase or decrease stability and how far a tree cover, compared with a shrub cover, increases instability (R. J. Eyles, Victoria University, pers. comm.). It is evident that considerable survey work will have to be undertaken as less desirable hill sites are brought into use, for no simple general rules apply in an area of extremely variable soil, rock and slope conditions.

Vegetation

The effect of vegetation in protecting slopes against wash, rill and gully erosion has been well established by experiments carried out under the aegis of the United States Soil Conservation Service in the 1930s and by much experience elsewhere. The effect of tree cover upon slope stability has been the subject of greater debate. Arguments that tree cover increased creep rates and that instabilities were greater under forest cover conflicted with indications that trees improved slope stability. An analysis by Brown and Sheu (1975) showed that: tree roots provide mechanical reinforcement of the soil; vegetation provides a vertical slope surcharge; soil moisture content and water table levels are lowered by evapotranspiration from trees – all of these changes would increase slope stability. On the other hand wind acting against trees may cause surface shears; this process may be significant in places like Fiordland where soils are very shallow and slopes very steep, and it may be of importance in the Wellington area on steep slopes with thin soils.

The importance of tree removal for slope stability is obviously very great in New Zealand. Very large areas have been deforested in the last century and substantial areas have been, and are being, reafforested. Afforestation is also a major soil conservation measure. The quantitative significance of forest removal has been examined by O'Loughlin and Pearce (1976) for part of North Westland. They found that in areas underlain by massive Tertiary sandstone and siltstone denudation rates were increased 40 times after forest removal. Landslide densities increased from $1/km^2$ to $20/km^2$. Forest operations had a less severe effect upon denudation. The analysis showed that the supporting effect of the tree root networks on saturated soils is very high.

Similar analyses to those carried out by O'Loughlin and Pearce are now being undertaken, from Waikato University, upon slopes in the weathered greywacke of the Hapuakohe Ranges and on Cenozoic 'soft' sedimentary rocks of the King Country.

Instrumentation and Equipment

One of the difficulties which quantitative analyses of slope stability has only recently overcome is that many soils on slopes are in coarse grained colluvial materials which owe part of their apparent cohesion to the root network within them. Standard laboratory triaxial and shear box equipment is not capable of measuring shear strengths of very large undisturbed soil samples and cannot be used to study the effect of roots upon soil strength. The introduction, by O'Loughlin (1974), of the field shear box to New Zealand was, perhaps, the most important development in slope stability studies which are related to shallow landslides.

Two types of field shear box are currently in use; one open only at the base so that shearing forces are measured parallel to a presumed basal failure surface, and a second box open at the base and on two sides so that shearing forces are also measured laterally. The open sided box is most useful for work in soils with a dense root network. Both boxes and their ancillary equipment can be carried into inaccessible areas, although the use of high normal loads necessitates the transport of heavy lead or concrete weights. The open sided box currently in use at the University of Waikato is shown in Fig. 1.

Other forms of instrumentation in use in New Zealand are relatively standard. Measurements of soil creep by Owens (1969) in the Chilton Valley near Cass were most successful when plastic tubes were used. The other methods using T–bars, columns of stones and Young pits (see Selby 1966b) were less successful. Aluminium cones attached to wires reaching the surface of the ground were used by Selby (1968, 1974b). The measured rates in New Zealand are similar to those determined elsewhere in temperate regions.

Earthflows of various kinds have been measured by a variety of targets erected in the moving mass and also by targets attached to continuous chart recorders (R. K. Smith, Ministry of Works and Development, pers. comm.).

The very irregular occurrence of landslides in time and space militates against long-term observations. Most analyses are therefore of a 'post-mortem' nature and, hopefully, these will in future be more concerned with geotechnical information and with stability analyses than has been the situation in the past, when most reports were essentially descriptive.

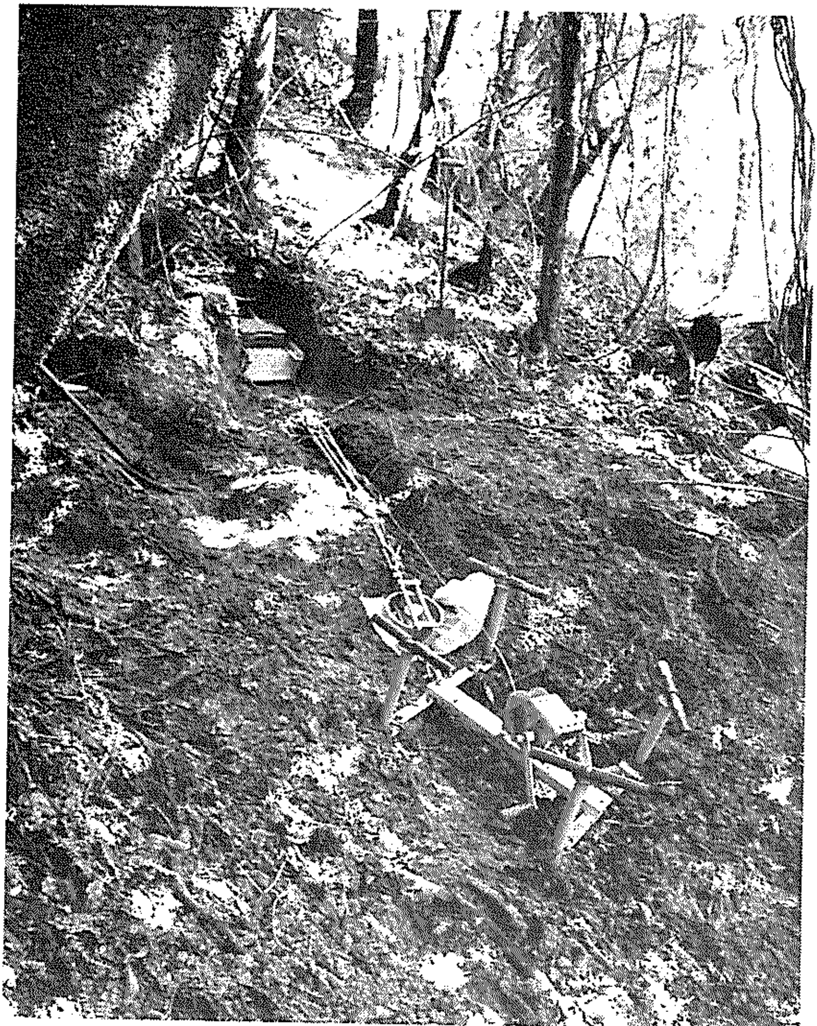

FIG.—1 An open side field shear box in use in a forested area with coarse soils. The force is applied by the winch and measured on the proving ring. The box is placed on an *in situ* excavated monolith and the normal load is applied by the weights on top of the box.

Morphometry

Although New Zealand has a great range of landslide types and modes of occurrence relatively few detailed studies of their morphometry have been made. The studies of Blong (1971, 1973, 1974) and Crozier (1973) are notable attempts to consider the relationship between landslide forms and processes. Blong found that there is little discernible relationship between hillslope form and landslide form and that there is only a limited connection between landslide size and landslide shape. Attempts to distinguish between debris slide–debris avalanche–debris flow types of translational slides using numerical taxonomic techniques were not successful. Crozier, by contrast, found that his seven morphometric indices were good indicators of the processes responsible for producing variation in landslide form.

Prehistoric Landslides

The destruction of the forest cover of New Zealand has presumably happened repeatedly in the past as a result of Pleistocene climatic changes and, in the last 1000 years, as a result of human interference. There is increasing evidence that the northern part of the North Island remained largely forest covered even during glacial maxima (A. T. Wilson, Waikato University, pers. comm.) but much of the remainder of the country suffered repeated vegetation changes in response to climate. It is probable, therefore, that many ancient landslides may be related to associated vegetation changes, as well as to tectonic instabilities and other causes.

In the northeastern part of the North Island ancient phases of instability are recognisable because instability destroys the mantle of volcanic ash beds over the landscape. Thus the youngest complete ash bed covering a surface provides a datum indicating stability since it was deposited, and older phases of denudation may be interpreted from the presence and absence of ash beds. Ash beds have been used by Pullar (1965, 1970), Grant (1963, 1965), Selby (1966a) and Gage (pers. comm.) to date phases of landsliding and other forms of erosion.

Gage working in the inland Gisborne region has been able to identify areas which have been undisturbed for 30,000 years; limited areas above 1000 m have been affected by solifluction and scree formation during a time of lowered tree line between 11,000 and 30,000 years ago; a general period of slope undercutting and instability was presumably caused by tectonic uplift and consequent stream incision after 11,000 years ago; and other areas are affected by deforestation in the last 100 years.

Pullar has been able to determine that the large landslide whose debris impounded Lake Waikaremoana occurred between eruptions of Taupo Pumice and Waimihia Lapilli, that is between about 1800 and 3500 years ago. He has also dated periods of deposition in the Gisborne lowlands and Galatea Basin which imply erosion in the uplands. Selby dated deep rotational slumps near Whitehall as occurring between 1800 and about 13,000 years ago at a time when the area was fully forested.

The most comprehensive pattern of erosional events in late Holocene time which has been elucidated is that described for the Hawke's Bay region by Grant. He discovered that there have been a number of erosional intervals since the 13th century including the Waihirere and Matawhero episodes which closed about the end of the 14th and 16th centuries respectively, and an early 19th and late 19th century phase. These early episodes and the modern erosional phase appear to be related to periods of storminess, and the increasing size of the particles in valley floor aggradations implies that landsliding and gullying are cutting into progressively less weathered bedrock.

In other parts of the country Scott (1963) has recognised past phases of erosion from paleosols in fan deposits of the Godley Valley, and Burrows (1975) has dated a landslide in the upper Acheron Valley at 500 ± 69 by radiocarbon. It is thought that this and other landslides in the area have been triggered by earthquakes.

Relict in the landscape of central and southern New Zealand are many indications of past erosional phases and landslides attributable to periods of colder climate. Cotton and Te Punga (1955a, b) and Stevens (1957) have described solifluction debris from the Wellington district and McCraw (1965) has described periglacial phenomena in central Otago. All of these phenomena indicate periods of rapid landscape development in response to past climatic conditions, followed by stabilisation of deposits in climates more favourable to plant growth.

Magnitude and Frequency of Landsliding

Nearly all examples of regional landsliding in recent years have occurred during individual storms or as a result of prolonged wet periods. The frequency with which such events are repeated is thus of considerable concern. Unfortunately the length of continuous daily rainfall records for much of the country is relatively short. The return period for storms can only be calculated where the length of record is equal to or greater than the recurrence interval. The method frequently used is that of Dalrymple (1960) in which:

$$\text{return period} = \frac{N + 1}{M}$$

where N equals the number of years of record and M is the rank of the individual item in an array. The frequency of very long return period storms can thus not be assessed reliably.

Using this method it has been found that storms with a return period of as little as two years have caused landslides where soil and vegetation conditions are near a critical condition of instability. More commonly trigger events have a return period in excess of 10 years. The finding by Selby (1974a) that dominant land-forming events in the greywacke hills of the Auckland Province are rainstorms which for pasture areas have a return period of 30 years, but in forest areas are of such magnitude that they have a return period of 100 years, helps to put the effects of landuse upon landsliding into perspective.

Rates of Denudation

Determining the frequency of landslide-producing storms has the additional value that it permits a calculation of the denudation rate, as the total rate is the sum of that for storms of each magnitude. Calculation of average volumes of landslide scars and of areas affected is relatively easy and the calculated denudation rates are more easily derived than those by most other methods, and are at least as accurate.

TABLE—1 Denudation rates produced by landslides alone

Soil or rock	*Vegetation*	*Locality*	*Surface lowering (mm/year)*	*Source*
weathered greywacke & colluvial volcanic ash	pasture	Hapuakohe Ra, Waikato hills	1.0	Selby, 1976b
weathered greywacke & colluvial volcanic ash	forest	Hapuakohe Ra Waikato hills	0.25-0.5	Selby, 1976b
loess and volcanic ash on siltstone	pasture	Tangoio, Hawke's Bay	1.5-2.3	Eyles, 1971
shattered greywacke	forest and shrubs	Southeast Ruahines	0.8-5.2	Mosley, 1977
weathered greywacke	forest	Stokes Valley Wellington	0.03	McConchie, 1977
gravels & sandstone	forest	North Westland	0.1	O'Loughlin and Pearce, 1976
gravels (Waimaungan)	clear-felled young pines	North Westland	1.1	O'Loughlin and Pearce, 1976
weathered gravels (Old Man)	clear-felled young pines	North Westland	1.4	O'Loughlin and Pearce, 1976
sandstone	clear-felled young pines	North Westland	4.0	O'Loughlin and Pearce, 1976

TABLE–2 Denudation rates in New Zealand (collated by Mosley, 1977)

Locality	*Surface lowering (mm/year)*	*Source*
Tukituki Valley (within N.Ruahines)	1.5	de Leon, 1976
Lake Tutira	1.7 (minimum)	Grant, 1963
Mangahao Valley	1.6	Thompson, 1976
Waipaoa Valley	6.5	Thompson, 1976
Torlesse Stream	0.02	Hayward & Blakely, 1976
Roxburgh Lake	0.25	Thompson, 1976
Frazer Dam.	0.03	Thompson, 1976
Tengawai Valley	0.2	Cuff, 1974
Opihi Valley	0.08	Cuff, 1974
Opuho Valley	0.06	Cuff, 1974
Otaki Valley	1.0	Manawatu Catchment Board
Shotover Valley	0.8-2.3	Ministry of Works and Development

c.f. Himalayas 1.46 (Curray & Moore. 1971); European Alps 0.8–2.0 (Clark & Jager, 1970); Taiwan Mountains 10.7 (Li, 1976); World Steeplands 1.0 $\pm$0.8 (Young, 1969).

Note: a surface lowering of 1mm/year $\simeq$ removal of 10 m^3/ha/year = m^3/km^2/year

In Tables 1 and 2 are shown all known data on the denudation rates in New Zealand for landsliding alone and from all processes. It can be seen that for landsliding alone New Zealand rates are close to those given as a world steeplands average (Young, 1969; Selby 1974c). Exceptionally high rates occur in the unstable mudstone lithologies of inland Gisborne and locally where clear felling and adverse lithologies occur together on the West Coast. No reliable data exist for the high alpine areas in greywacke lithologies but these are likely to be well above the rates which have been measured. The long term high rates in the Ruahine Ranges (Mosley, 1977) indicate that, although human interference can accelerate natural rates for a while, they do not greatly exceed the 'natural' geological rate for long.

CONCLUSION: PROBLEMS AND TRENDS

Problems

One of the major deficiencies in slope stability studies in New Zealand has been the lack of agreement on terminology and presentation of data.

Several classifications are commonly in use: that by Campbell (1951) was partly inspired by the work of Sharpe (1938) and intended for local application. It has been found very useful but classifications are most valuable when they enable comparisons to be made with data from elsewhere. The work of Varnes (1958) has also been used, but this is now largely superseded by that of Hutchinson (1968). It is suggested by Northey *et al.*, (1975) that this trend be followed and Hutchinson's classification be used.

A second deficiency is the lack of uniformity in the terminology applied to individual landslides. Crozier's (1973) terms have been found useful both in New Zealand and overseas and it is suggested that they should be followed. In Fig. 2 his terms, and measurements used in morphometric indices, are indicated and to these have been added suggestions for measurements to determine landslide scar volumes for stability analyses.

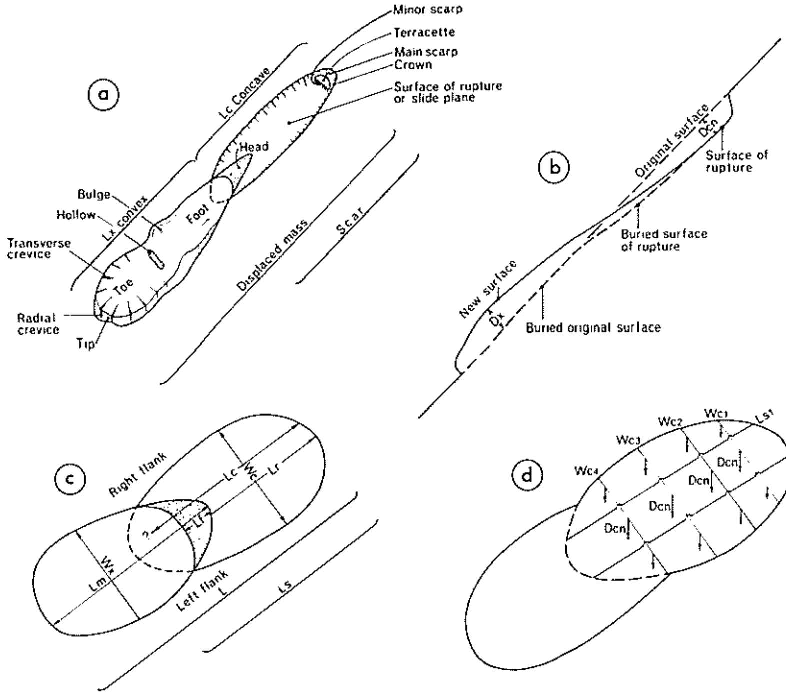

FIG.—2 Suggested nomenclature of landslides and morphometric indices. After Crozier (1973) with additions.

A third problem is that few reporters of landslides use similar descriptors of features and precipitation. Rapp (1974) has suggested the format which is shown in table 3 and wide use of this would enable landslide-producing storm events to be compared on a world basis.

Trends

Work in progress reported by correspondents at Canterbury and Victoria Universities indicates that much interest is being focussed on the economic significance of landsliding as it affects transport routes and urban areas. The Forest Research Institute and Waikato University are concentrating on slope stability analyses in relation to deforestation and afforestation by using field shear boxes. At Auckland University a number of projects are underway in which lithological and structural influences on stability are being analysed.

The general trend is for less descriptive work and more quantitative work to be undertaken, with a much closer attention being paid to the nature of the rock and soil materials involved in landslides. These healthy trends indicate that geotechnical studies are now replacing the pioneer descriptions as many of the direct relationships between landform change, climate and vegetation are now recognised. The new trend will require greater technical competence and the use of modern analytical equipment, but it should lead to a greater understanding of the processes and nature of landsliding and the material it involves.

TABLE—3 Data on extreme rainstorms and mass movements for two areas in the Hapuakohe Range (from Selby, 1976b)

		Mangawhara Valley	*Matahuru and Mangapiko Valleys*
1.	Latitude, Longitude	37°S, 175°E	37°S, 175°E
2.	Altitude (m a.s.l.)	150-350	90-420
3.	Mean annual temperature (°C)	14	14
4.	Annual precipitation (mm)	1800	1800
5.	Rainfall records since	1956	1970
6.	Relief range (m)	200	330
7.	Bedrock	greywacke (sandstone and siltstone)	
8.	Regolith type, depth (m)	In situ weathered greywacke with colluvial volcanic ash, 5-20	
9.	Vegetation	pasture grasses	pasture grasses
10.	Rainstorm date	28.2.1966	25.12.1973
11.	Rainstorm total (mm)	150-230	190
12.	Rainstorm duration (hours)	24	15
13.	Rainstorm type	cyclonic with local orographic lifting	local thunderstorm
14.	Rainstorm area[1] (km^2)	250	2.8
15.	Rainstorm return period	10 or >10	>20
16.	Slope erosion type[2]	debris avalanche and debris slide	
17.	Slope gradient (degrees)	32	34
18.	Depth of scars (m)	0.92	0.9
19.	Width of scars (m)	12	12
20.	Denudation/catchment area[3] (mm/km^2)	80 mm/0.5 km^2 to 10mm/20 km^2	40mm/2.8 km^2
21.	Mass transport form	debris flow	debris flow
22.	Mass deposition form	colluvium reworked in stream bed	
23.	Mass max. depth (m)	1.5	1.8
24.	Mass gradient (degrees)	5-25	5-25

Notes. [1] Area in which mass movement occurred
[2] Classification of Varnes (1958)
[3] Vertical downwearing of the landscape
Denudation rate is calculated from volume of erosion scars.

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RIVER FLOW MEASUREMENT

J. R. Waugh and J. K. Fenwick

ABSTRACT

The paper reviews river flow measurement as it is currently practised in New Zealand. The first section deals with the use of calibrated flow measuring structures. This is followed by a detailed examination of velocity area methods of stream gauging, with emphasis on current meter gauging.

A number of new techniques have been developed overseas, these include the moving boat method, ultrasonic and electromagnetic flow measurement. These techniques are discussed briefly as they are not used in New Zealand at present. Chemical gauging and related techniques are also discussed in this section.

The final section deals with recording of water–level and the tower installations used for this purpose in New Zealand. Throughout the paper commentary is provided on the accuracy of the techniques and on operational field problems encountered in New Zealand.

INTRODUCTION

This paper reviews 'river flow measurement' in the broadest sense as it is currently practised in New Zealand. Commentary on the applicability of particular methods of streamflow measurement is combined with notes on accuracy and discussion of new methods not yet in use in New Zealand.

'River flow measurement' can be interpreted to mean the whole process of recording water–levels, streamflow measurement and computation of discharge values to produce a usable record of river flow. This last activity is dealt with in the paper by Dr R. P. Ibbitt in this volume.

As no proven technique for continuous monitoring of discharge of natural channels exists, it is customary to measure stage continuously, then relate discrete measurements of discharge to the stage record to produce continuous discharges (Halliday and Chapman, 1976).

In New Zealand the measurement of river discharge at a cross section on a river is commonly achieved by one of three techniques: calibrated structures; velocity–area gauging; chemical or dilution gauging.

CALIBRATED STRUCTURES

Structures are generally most suited to small streams, up to 30 metres in width. Provided the structures are maintained to meet their design requirements, they permit flows to be measured accurately over long periods of time, which makes them particularly useful in experimental or research work. Calibrated structures are virtually essential in small urban catchments where flood runoff is so rapid that rating of a natural stream station by conventional current meter gauging is exceptionally difficult.

When a structure is being built, it is good practice to conform to the standard design for that structure. This allows immediate use of the standard stage–discharge rating to compute discharge values and also means that a minimum of check gauging is required to confirm the validity of the standard rating. Neglecting to do so or installing non–standard structures negates this most useful benefit such that one is then faced either with the sometimes difficult task of field rating the structure or with a costly model rating under laboratory conditions. De Laine (1964) describes a method for field calibration of structures.

Field experience in New Zealand has served to emphasise the importance of stream channel conditions, both upstream approach conditions and downstream tailwater conditions, on flow measuring structures. Standard texts dealing with various types of structure tend to emphasise the important physical dimensions of the structure but offer little advice on site conditions.

H Flumes

H flumes have been installed on small catchments and runoff plots, for example, at Makara and Moutere Soil Conservation Reserves. Larger H flumes (up to 1.83 m) were installed at Pukewaenga and Pukeiti (Northland) and at Camp Stream. Flumes were also used in the central North Island at Purukohukohu and have recently been installed by the Grasslands Division, Department of Scientific and Industrial Research, at Ballantrae near Palmerston North.

Flumes are most useful where the stream transports reasonable amounts of sediment or floating debris. U.S. Department of Agriculture (1962) provides basic information on various types of flumes.

A problem encountered with larger flumes is their lack of sensitivity at very low flows. At several stations this has been overcome by adding a low–flow V–notch on a trough or box, below the flume outlet. In some cases the recording of low flow was separate, in others it was combined with the flume to record on a single water–level recorder. Field operation of these low–flow weir boxes generally has been satisfactory but processing the data to produce a single record for the station can be difficult.

Weirs

Sharp–crested weirs are generally considered the most accurate structures for measuring streamflow. Rothacher and Miner (1966) state that: "if properly maintained, they can give measurements accurate to within 1 or 2 percent". Flow measurements of this accuracy are dependent on very low approach velocities, a clean sharp–crested weir plate, and a fully ventilated nappe. To minimise rating errors at very low flows, Rothacher and Miner (loc. cit.) recommend that the head of water over a sharp–crested weir should not be less than 0.06 m unless careful rating measurements are made on the specific weir being used. They also quote two examples from well operated small catchment research areas in the United States, where the values of flow measured "are in error by not more than 3 to 5 percent" in one case and have "an error of about 3 percent of annual flow" in the second example. They conclude that "these examples probably approach the best accuracy obtainable under practical field conditions".

A feature of weir construction in New Zealand has been the use of a standard design for a group of catchments. This simplifies construction, reduces costs, should result in standard ratings, and produces flow data of comparable accuracy for all catchments in the group.

In siting weirs in a catchment particular attention needs to be paid to both approach conditions and downstream conditions. Approach velocities should be kept well below 0.3 m/s to minimise error at low head (Kulin and Compton, 1975, Table 4.1). However, this may not be possible during flood events. Drowning of a weir by backwater from downstream during floods can affect an otherwise satisfactory site. This condition may be difficult to detect prior to construction. Siting weirs on bedrock outcrops above waterfalls or at the head of steep rapids will generally overcome these problems and at the same time minimise the chance of leakage around a weir. Weirs should not be used where excessive amounts of suspended sediment, bedload and floating debris are encountered. Weirs are unsatisfactory in these conditions as the weir pond traps sediment thus altering the weir rating, while debris can clog the V–notch producing grossly inaccurate flow measurements. In these conditions a broad crested triangular weir or a Crump weir may be more appropriate.

Compound weirs

Compound weirs, consisting of a 0.61 m 90° V–notch with sloping wingwalls (usually at 1 in 5 slope), were built at a number of stations during the International Hydrological Decade Representative Basin Programme 1965–74 (Ministry of Works and Development, 1965). Fig. 1 shows a typical large compound weir in Northland, New Zealand. Compound weirs have been installed in 12 catchments at the Moutere Soil Conservation Reserve replacing the earlier H flumes. Compound weirs give accurate flow measurements, particularly for low flows and also allow a wide range of discharge to be measured, for example 0 to 100 m^3/s for some of the larger weirs in Northland. Flow data

FIG. — 1 Compound weir, Northland, New Zealand.

from small catchments in Northland, up to 12 km², show that for over 95% of time flows are within the V–notch section of the compound weirs, that is, in the section where flow measurement is most accurate and where there is also good sensitivity, sensitivity being defined as the change of stage per unit change in discharge.

Reinhart and Pierce (1964) recommend the use of 90° V–notch weirs where discharge 95% of the time is expected to be between 0.28 and 226 l/s, and 120° V–notch weirs where flow 95% of the time is expected to be between 2.8 and 566 l/s. The use of debris screens to prevent blockage of the notch by grass, leaves or twigs is also recommended (Reinhart and Pierce, 1964; Gleeson, 1961). 120° V–notch weirs are less prone to blocking by floating debris.

90°V–notch weirs

These have been constructed on research catchments near Reefton in 1974 and at Donald Creek in the Nelson area in 1975 by New Zealand Forest Service. Between 1955 and 1963 similar weirs were established on research catchments at Taita by Soil Bureau of D.S.I.R. 90° V–notch weirs set into horizontal broad-crested structures have been built in regional catchments in Otago (Gimmerburn and Nobles Stream) by the Water and Soil Division of the Ministry of Works and Development.

120°V–notch weirs

In 1976-77 120° V–notch weirs were built in four catchments at Golden Downs near Nelson (Fig. 2) by the Water and Soil Division of Ministry of Works and Development.

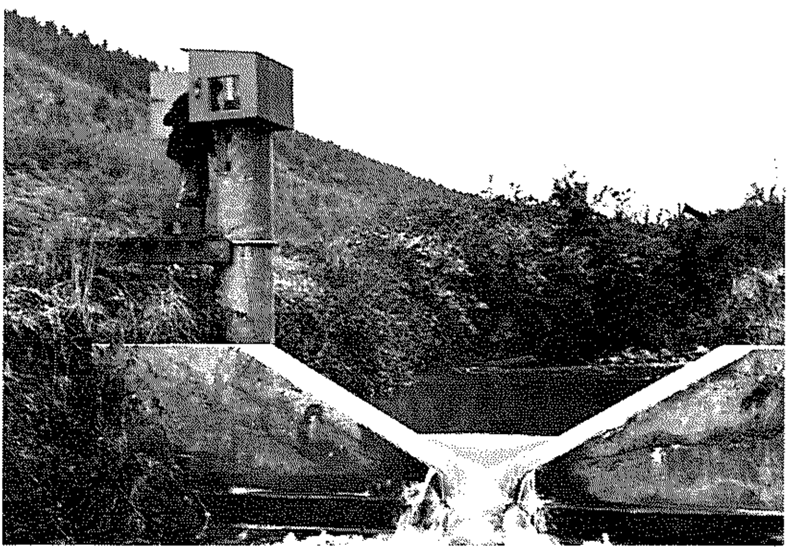

FIG.—2 120° V–notch weir in Nelson District, New Zealand.

Crump weirs

Two Crump weirs were built in New Zealand during 1975: one near Whangarei (Fig. 3) and the other on a small urban catchment in Rotorua. These flow measuring structures can be adapted to a wide variety of site conditions, can measure a large range of flows, and continue to operate in a satisfactory manner with a moderate sediment load. A variety of compound structures can be built to provide sensitivity for low flow measurement together with the capacity to handle large sediment laden flood flows. To date the two Crump weirs built in New Zealand have operated in a satisfactory manner.

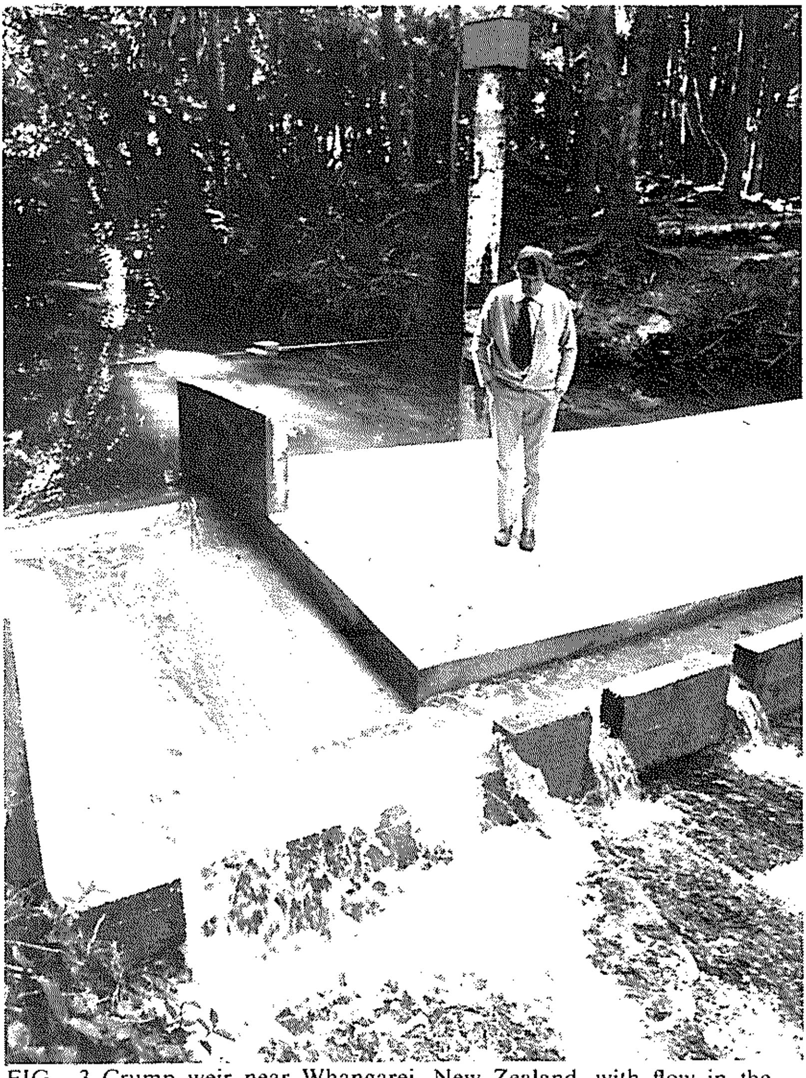

FIG.—3 Crump weir near Whangarei, New Zealand, with flow in the low–flow section only.

Crump weirs are comparatively simple to construct and have been shown to be relatively tolerant of poor site conditions (White, 1975).

The Crump weir has a simple monotonic stage–discharge relationship whilst operating up to the total head at which tailwater reaches 75% of that upstream (relative to crest height). Using crest tappings designed by Crump it is possible by dual measurements to get reasonable results up to 90% submergence (Twort *et al.* 1974).

Standard ratings are available for Crump weirs, and White (1975) has shown that there is little value in field rating a carefully constructed Crump weir.

Other Structures and Bed-controls

A wide variety of weirs and flumes have been used in New Zealand, including a number of 'one off' non–standard designs. For example at some stations, often on larger streams (e.g., Selwyn catchment 164 km^2), a bed–control has been constructed, to give a fixed cross section for water–level recording, in an otherwise mobile bed. These stations are then rated by use of conventional gauging methods. However, because experience has shown that using non–standard designs can lead to problems in construction, operation and field ratings there are great advantages in using one of the numerous standard weirs or flumes now available.

Cost-effectiveness of Structures

Organisations and individuals involved in hydrological research and streamflow measurements often react adversely to the use of flow measuring structures because of the high initial capital cost. However, when one considers the cost of establishing, field rating and operating a 'natural' stream or river station, the situation is put in perspective.

It is estimated that for an accessible site, without special construction or gauging difficulties, the cost of establishing a water–level recorder station, developing and maintaining ratings, servicing the recorder and processing the data for two years is $10,000. To operate the station for 5 years would probably cost an additional $5,000, or a total of $15,000, and for 10 years a total cost of $20,000 is estimated (1977 costs). These estimates assume a high frequency of gauging in the first two years to establish a stage–discharge relationship, followed by fewer gaugings in subsequent years, to maintain the rating. Where floods produce rating changes more gaugings would be required to re-establish the rating.

Weirs or flumes can be constructed for sums ranging from $400 up to say $10,000 for a large (100 m^3/s) compound weir or Crump weir. The major advantages gained using precalibrated structures are:

(a) operation costs are lower as little if any gauging is required;
(b) there is a fixed cross-section hence no rating changes occur;
(c) more accurate data are obtained throughout the entire period of record;
(d) flow data are immediately available if a standard calibrated structure is built;
(e) apart from routine servicing of the water–level recorder commitment of field staff is minimal.

We conclude that more use should be made of precalibrated flow measuring structures in New Zealand, particularly where the station is on a small stream and will be operated for more than five years.

VELOCITY–AREA STREAMFLOW GAUGING TECHNIQUES

The conventional current meter gauging methods of streamflow measurement and a number of relatively new gauging techniques are all variations of the velocity–area approach to streamflow measurement. This section examines the various techniques available and also briefly considers subsidiary methods for measuring very small flows and flood flows.

Current Meter Gauging Methods

The first current meter gaugings in New Zealand were carried out around 1903. Although the instruments have improved, there has been relatively little change in the use of current meter gauging because it is widely applicable.

Alternative techniques generally have more specialised applications or are useful in particular conditions. Most current meter gauging in New Zealand has been carried out by staff of Ministry of Works and Development and the catchment authorities. Flows measured range from minute low flows under drought conditions, to 3740 m^3/s on the Grey River at Dobson.

The method of measuring stream discharge by current meter is to measure velocity and depth at suitably spaced points across a section of the stream. For each partial section the cross sectional area is computed and multiplied by its measured mean velocity to give its discharge. The summation of all the discharges for the partial sections is the total discharge of the stream.

Current meters

Current meters in use in New Zealand are of two main types: vertical axis and horizontal axis. Vertical axis meters in common use are the Price AA (Gurley), Pygmy, and less often the Watts. Horizontal axis meters in use are the large model Ott, the small Ott, and the Amsler.

Depth measuring equipment

Depth measuring and meter suspension equipment are normally combined, and are of two main types: wading rods and a cable–weight assembly. A rod, 13 mm in diameter and graduated at 10 mm intervals, is used for measurement at sites where wading is possible, with the depth being read directly from the rod and the meter being positioned on the rod by a sliding clamp. Cable suspension with the meter mounted above a streamlined Columbus weight is used when gauging larger streams from bridges, cableways, boats, or slackline cableways. A specially designed winch controls the cable, and has a counter for depth measurement and meter positioning and the facility for connecting the meter–contact/headphone circuit.

Width measuring equipment

With wading and boat gaugings, a graduated tape or steel tagline is stretched across the measurement section and the meter positioned along this tape. At permanent structures, distances are pre-marked.

Selection of gauging sites

A gauging site should be chosen on the stream firstly to ensure that its location suits the purpose of the data collection. Although preferred, close proximity to the water–level recording station or site of interest is not imperative. Should there be any inflows to or outflows from the stream between the gauging site and site of interest, however, these must be measured or otherwise taken into account for the data to be usefully applied. Where losses or gains through subsurface gravels are known or suspected, the gauging site should be as close as practical to the site of interest. Underflow of water through gravel is a widespread problem in New Zealand, particularly in the South Island. To minimise the problem it is common practice to select gauging stations in rockbound gorges where most of the water leaving the catchment will be forced to the surface.

The importance of careful site selection cannot be over emphasised; it is a critical factor in the accurate measurement of streamflow.

Velocity measurement

A current meter measures velocity at a point, but the method of making discharge measurements at a cross section requires determination of the mean velocity at each vertical. This can be done by measuring velocities either at many points in the vertical or at a few points known to give a close approximation of the mean velocity. Several different methods can be employed.

The vertical–velocity curve or multi–point method involves measurement of velocities at many points in the vertical (usually every one-tenth of depth) and a depth–velocity curve is drawn.

The vertical–velocity curve method is not generally used for routine measurements, mainly because of the extra time required to carry out the gauging. It is recommended that it be used once or twice at each new permanent gauging site to confirm that more approximate methods are satisfactory. It can also be used to determine any coefficients required for incomplete measuring methods and to examine the flow characteristics of a particular site. It is particularly valuable as a means of determining flow characteristics at a new gauging station.

The two–point method (0.2 and 0.8 depth) is widely used in New Zealand, U.S.A. and Canada. It is reported by Carter and Anderson (1963) and Buchanan and Somers (1969) that the 0.2 and 0.8 depth method has virtually become the standard international method of current meter gauging. Buchanan and Somers (1969) state:

> This method is based on many studies of actual observation and on mathematical theory. Experience has shown that this method gives more consistent and accurate results than any of the other methods except the vertical–velocity curve method.

In New Zealand the two–point method is used in routine gauging work wherever possible. The main limitation to this is one of boundary effects both from the streambed and the water surface, on the current meter performance. It is U.S. Geological Survey general procedure that this method be used at all times except at depths less than 0.76 m because the current meter comes too close to both the streambed and the water surface to give dependable results.

Pygmy and small Ott meters are less affected close to the surface and bed because of their considerably smaller size. The use of these meters in a reasonably laminar flow over a relatively smooth streambed would allow the use of this method to 0.5 m depths. Below this depth, the 0.6 depth method must be used.

Table 1 (S. E. Rantz, U.S. Geological Survey, pers. comm.) sets out the depth limitations in relation to the method of velocity observation.

TABLE 1 – Current meter and velocity–measurement method for various depths (wading measurement).

Depth (*ft*)	*Depth* (*m*)	*Meter*	*Velocity Method*
2.5 or more	0.76 or more	Type AA (or type A)	0.2 and 0.8
1.5 - 2.5	0.46 - 0.76	As above	0.6
0.3 - 1.5	0.09 - 0.46	Pygmy[1]	0.6

[1]Used when velocities are less than 2.5 ft/s (0.76 m/s).

The 0.6 depth method, as with the 0.2 and 0.8 depth method, has been widely tested and is found to give reliable results. It should be used as an alternative to the 0.2 and 0.8 depth method where the depth is too shallow (below 0.76 m) or the meter is placed at a distance above the sounding weight which makes it impossible to place the meter at the 0.8 depth. This method can be used satisfactorily down to about 0.09 m depth by employing a Pygmy or small Ott meter. In conditions of rapidly changing stage, where a measurement must be carried out quickly this method can be used to save time.

The three–point method consists of observing the velocity at 0.2, 0.6 and 0.8 of the depth, thereby combining the two–point and 0.6 depth methods. The mean velocity is computed by averaging the 0.2 and 0.8 depth observations and then averaging the result with the 0.6 depth observation. If more weight to the 0.2 and 0.8 depth observations is desired, the arithmetic mean of the three observations may be used. However, the first procedure is usually followed.

The four–point method (1–4–6–9) has been used to some extent in this country but is apparently not used by the U.S. Geological Survey. It appears to have little advantage being probably only slightly better than the 0.2 and 0.8 depth method, while taking almost twice as long. It may occasionally be of use, however, as a short cut to the multi–point method for some purposes.

The 0.2 depth method consists of observing the velocity at 0.2 of the depth below the surface and applying a coefficient to this to obtain the mean in the vertical. It is used by U.S. Geological Survey for gauging floods where the discharge in it, but this is very seldom accomplished when 25 verticals are used. Equal widths of partial sections across the entire cross section are not recommended unless the discharge is well distributed. Across the gauging cross section the width of the partial sections should be varied so that the discharge in each is about 5% of the total discharge.

After the meter depth is calculated and the meter positioned, it must be permitted to adjust to the current before starting the velocity observation. When adjusted the number of revolutions made by the rotor during a period of 40–70 seconds is counted.

The internationally acceptable standard gauging consists of velocity observations at 0.2 and 0.8 depths, on more than 20 verticals, with velocities observed for more than 40 seconds. If a large number of gaugings are made using this method, two-thirds of the measurements will have an error of less than 3.0% (Carter and Anderson, 1963).

In discussing the accuracy of flow measurements obtained at natural control sections, Rothacher and Miner (1966) state that at best, error might be controlled to ± 5%. In most cases estimates of streamflow made at natural sections would not be this good.

The main factor affecting the accuracy of current meter gaugings is the number of verticals at which velocities are observed. Kulin and Compton (1975) in discussing stream gauging errors give an example of a guaging with 10 verticals using 0.2 and 0.8 depth observations over 40 second periods. The standard error for this type of gauging was computed to be ± 4.4%. They state that 95% of measured discharges will be within ± 8.8%. If 50 sections were used in gauging, the overall standard error is reduced to ± 1.6%. In practice there is little value in using more than 30 verticals.

A feature of many rivers is the unstable nature of their gravel beds (Fig. 4), particularly those rivers on both flanks of the axial ranges in the North and South Islands. Experience in New Zealand has shown that floods will frequently alter the stage–discharge relationship at stations with unstable bed conditions. At some stations it is necessary to gauge the river as often as once a week in order to maintain an adequate set of rating curves. On smaller rivers bed–controls have been used to provide a stable measuring section. On large rivers with a great depth of mobile gravel solutions of this type have not been attempted.

It should be noted that the water–level recorder stations on these unstable rivers are the best sites available and frequently have rock-bound banks with a mobile gravel streambed. At worst, both the banks and bed are gravel and the river flows in 'braids' across a very wide river bed. On the Wilberforce River it was necessary to establish three recorder stations over a period of two years and two of these stations had to be abandoned when the river changed course during a flood. River flow measurement under these conditions is a very difficult task.

Jet-Boat Gauging

This method of streamflow measurement is also known as the 'unanchored boat' method. The method has been widely used in New Zealand and is particularly valuable for gauging larger rivers at difficult or inaccessible sites. In practice a tag–line is fixed across the river. Because of problems of stringing the tag–line the maximum span is around 100 metres. The highly manoeuverable jet–boat is then held in fixed positions under the tag–line, by balancing the water-jet thrust of the boat against the river flow. At each position or vertical, velocities are too great to obtain soundings or to place the meter at the 0.8 or the 0.6 depth.

This method is not in general use in New Zealand, but it could be usefully employed where adverse conditions preclude the use of the 0.6 depth, and 0.2 and 0.8 depth methods.

During periods of rapidly changing stage, gaugings should be made as quickly as possible to keep the change in stage to a minimum. This speed will not only give greater accuracy in ascertaining a mean gauge height, but will minimise errors caused by the shifting of flow patterns as the stage changes.

Steps to speed up a measurement are: use the 0.6 depth method; reduce the velocity observation time to about 25 seconds; reduce the number of verticals taken to about 15–18.

By incorporating all three of these practices a measurement can be made in 15–20 minutes. Carter and Anderson (1963) have shown that the standard error for a 25 second period of observation and using the 0.6 depth method of velocity observations with 16 verticals is 4.2%. The error caused by using the short cut method is generally less than the error that can be expected by the shifting of flow patterns during periods of rapidly changing stage.

Another technique used in New Zealand to obtain a series of measurements over a flood peak of short duration, is as follows:

(a) Take 10–15 verticals depending on the time available.
(b) Take velocity observations at 0.6 depth.
(c) Repeat velocity and depth observations at the same verticals, noting the corresponding stage heights at least four times throughout the period of the flood wave.
(d) Develop stage–velocity and stage–depth curves for each vertical site.
(e) Compute the discharge corresponding to selected stages by derivation of the partial discharges from the curves thus defined.

This is similar to the technique used in tidal gaugings.

Computation methods

Two slightly different methods are used to calculate current meter gaugings. They are termed the mid–section and mean–section methods. The latter has been adopted by almost all authorities in this country.

With the mean–section method, partial discharges are computed for partial sections between successive verticals. The velocities and depths at successive verticals are each averaged, and the section extends laterally from one observation point to the next. Section discharge is the product of the average of two velocities, the average of the two depths, and the distance between the two verticals.

Factors affecting accuracy

The number and placement of verticals in the measurement section is critical in the accuracy of a gauging. Generally, about 25–30 verticals should be used, although this number may be reduced slightly with a smooth cross section and an even velocity distribution. The main criterion for vertical spacing is that no partial section should have more than 10% of the total discharge in it. The ideal measurement is one in which no partial section has more than 5% of the total

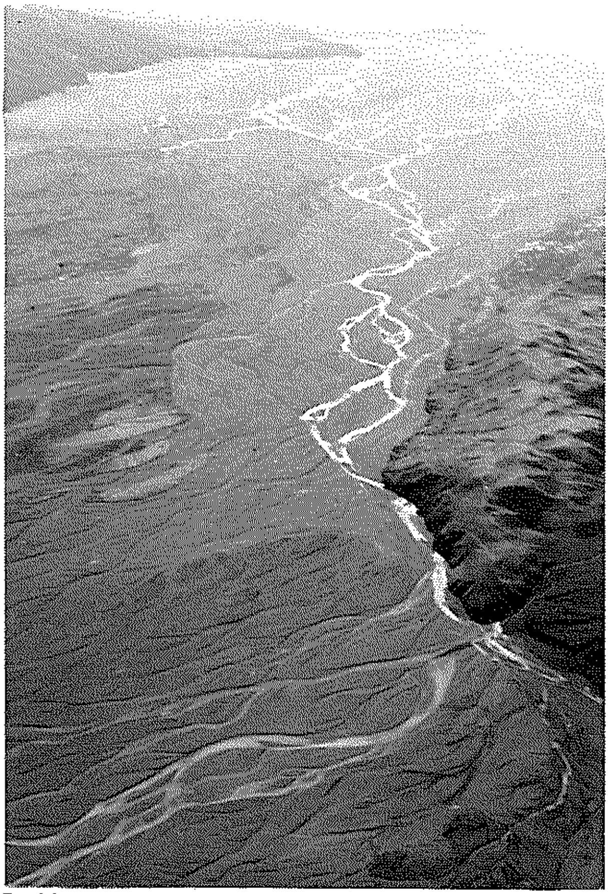

FIG.4 – Braided, unstable gravel–bed river in Canterbury, New Zealand.

conventional current meter velocity observations are made, using a crane mounted current meter over the bow. This gauging method should not be used in turbulent water as vertical motion of the boat will lead to inaccurate velocity observations, particularly if a vertical axis type of current meter (e.g. Gurley) is being used.

In Canada, the method has been used on rivers over 500 m wide, by using a Tellurometer with shore targets to provide distance measurements. In some cases a good quality echo sounder is also used to provide cross sectional data (Halliday and Chapman, 1976).

Moving Boat Method

The moving boat method was developed by the U.S. Geological Survey and has been described by Smoot and Novak (1968). To date the method has not been used in New Zealand as experience in Canada and USA indicates that the method is best used in large rivers over 500 m wide and in tidal estuaries or similar situations. It is also useful where stage is changing rapidly or the river is debris laden.

Buchanan and Somers (1969) state that "measurements obtained by the moving boat technique compare within 5% of measurements obtained by conventional means". Major advantages of the moving boat method are that it is quick to carry out and requires no fixed installations onshore.

Velocity–Azimuth–Depth Method

In tidal situations the direction of flow may vary with depth. To overcome this problem equipment that measures both velocity and direction is used. The U.S. Geological Survey developed a special velocity–azimuth–depth-assembly (VADA) for this purpose. In New Zealand use has been made of conventional sounding weights with Braystroke directional current meters.

Acoustic Velocity Meter

The acoustic or ultrasonic method of streamflow measurement uses the principle that the time required for a sound pulse to travel through a liquid is less in the direction of liquid movement, than against it.

Two principal systems exist at present; in one transducers are fixed in position and the station is calibrated by conventional current meter gauging, in another transducers move on a vertical assembly and the system is self calibrating.

Herschy (1974) states that "the ultrasonic method of river gauging will probably be most used at locations where no stable stage–discharge relation exists and in the lower reaches of rivers under tidal influence". He also states that "the accuracy of a single 15 minute measurement of discharge is estimated as ±4% at the 95% confidence level".

Ultrasonic flow measurement obviously has a place in the hydrologist's kit of tools. Its chief drawback is the high capital cost, current commercial models costing around $30,000. British costs in 1974 are quoted as £9,000 for equipment together with site engineering costs up to £15,000. The hydrologic data would have to be of considerable value to the community, or for a very high cost design project, to warrant the installation of ultrasonic gauging equipment. To date, this type of equipment has not been used in New Zealand.

Electromagnetic Method

As Newman (1876) explains "the main feature of this method is a large coil which is buried below the riverbed to generate a vertical magnetic field across the full width of the river...the control equipment drives current through the coil and measures the small potential, induced at probes in the river banks, generated by the water which is flowing through the magnetic field".

The accuracy of the electromagnetic method was checked at an experimental station under difficult flow conditions where the standard error of the mean for all runs was 1.7% at the 95% confidence level. This technique is still barely past the experimental stage and has not been used in New Zealand.

Integrated Float

The integrated float of streamflow measurement has been reported by Dyer (1970), who concludes "the results suggest that an accuracy of a few percent should be readily obtained with a properly engineered rising float system. The method involves releasing a float or series of floats from a device on the stream bed. The integral of stream flow is obtained from the distance travelled downstream by the object from the point of release to its emergence at the surface". The method does not require the river depth to be known, does not assume a particular velocity profile and gives greater accuracy in deeper rivers. John (1976) has reported on calibration of floats for this method.

Subsidiary Methods

Several complementary or subsidiary methods of streamflow measurement exist and are used in New Zealand when conventional current-meter gauging techniques are inappropriate.

Very small flows

As Buchanan and Somers (1969) state "the most accurate method of measuring small discharge is by observing the time required to fill a container of known capacity". The only equipment required is a stopwatch and a calibrated container. In the field it is often necessary to create a small earth dam to divert the flow through a pipe or length of roof guttering and thus into the container. Repeated measurements should be made and averaged to give a discharge.

Portable weirs and small Parshall flumes have been used in New Zealand to measure low flows. Using a 0.3 m 90° V–notch plate it is possible to measure flows from 0.6 l/s to 56 l/s within 3% accuracy.

Portable Parshall flumes can be used in shallow depths and low velocities. Under free flow conditions the accuracy of measurement is 2 - 3%. Construction plans and a rating for a modified Parshall flume are given in Buchanan and Somers (1969). This simple, lightweight, easily installed flume can be used to measure flows up to 14 l/s.

Velocity head rods, given suitable site conditions and careful use can be used to estimate small flows (Drost, 1963). Some use has been made of the method in New Zealand.

Flood flows

A short cut current meter gauging method was described earlier. Two other methods for measuring flood flows are the use of floats and an optical current meter.

Both surface floats and semi-submerged floats may be used to obtain an estimate of surface velocity during floods, in conditions where current–meters cannot be used. Buchanan and Somers (1969) recommend using a travel time greater than 20 seconds, measurement between two cross-sections, and the use of a number of floats distributed uniformly across the stream width. A coefficient of around 0.85 is commonly used to convert the surface velocity to mean velocity. Float measurements can give a flow measurement with an accuracy within 10% under good conditions and provided care is taken in making the observations.

The U.S. Geological Survey has developed an optical current meter which can be used to obtain surface velocities under flood conditions. The meter is a stroboscopic device designed to measure surface velocities in open channels, without any equipment being immersed in the stream. Smith (1961) has reported on the development and use of this portable hand held optical meter. It would seem to have some application in New Zealand, particularly for rivers with high velocities and large debris loads.

CHEMICAL GAUGING

The use of chemicals or tracers is another means of measuring river discharge. As the method does not involve area measurements it is particularly useful in rivers with mobile beds. The method involves injecting a known quantity of tracer at an upstream point and measuring the concentration at the downstream gauging station. Discharge can be computed by measuring the degree to which the tracer has been diluted. Two main methods are available:

(a) constant rate injection;
(b) injection of a 'slug' of material.

Accuracy of the discharge measurement depends on, among other factors, adequate mixing at the sampling station and an accurate concentration measurement. Determination of the mixing length required can also be a problem. Table 2 from Hill (unpublished) gives some examples of mixing lengths measured in New Zealand rivers. Day (1977) has also published results of research into mixing lengths for mountain streams in Canterbury.

TABLE 2 – Minimum lengths required on various streams of the Canterbury region, to mix injection solution of sodium dichromate

River	*Discharge litres/sec*	*Mixing lengths in metres*
Camp Stream	22.3	15
Reynolds	36.8	45
Garry	89.2	122
Garry	439	195
Ryton	705	365
Broken River	705	305
Broken River	881	457
Porter	1491	457
Potts	1880	549
S. Ashburton	2488	975
Potts	2602	732

The constant rate injection method involves the use of calibrated, controlled injection equipment. The Water and Soil Division, Ministry of Works and Development has shown that a 'Mariotte Vessel' is a satisfactory means of achieving a constant rate of injection under field conditions. The constant rate injection method allows the downstream sampling to be carried out for a limited period during the passage of the 'concentration plateau'. Both methods assume that no tracer is lost between the injection and sampling stations by either physical or chemical action. Rivers with areas of dead water or backwaters, which can trap the tracer, are not suitable for chemical gauging.

A slug injection, referred to as the 'gulp' method, is simple to apply at the injection point. The main disadvantage of this method is that the tracer slug has to be sampled over its entire time of passage, at some downstream point where mixing is complete.

The Agriculture Engineering Department at Lincoln College has carried out tests using a portable overhead sprinkler injection line system, which is strung across a river to allow dye injection across the full width of the stream.

A wide range of materials has been used for chemical gauging, including fluorescein, sodium chloride, sodium dichromate, rhodamine dyes and radioactive tracers. Rhodamine WT is now widely recommended as being the most satisfactory material for discharge measurements, or any other application requiring a high percentage recovery of dye (Wilson, 1968).

Accuracy of chemical gauging is potentially very high, provided the method is carefully used under suitable field conditions. The salt velocity method has a potential for better than 1% accuracy (Kulin and Compton, 1975) and Kilpatrick (1968) achieved slightly better than $\pm$ 2% with continuous injection in the laboratory. Kulin and Compton (1975) state: It can be assumed that similar accuracy could be attained in artificial conduits under good conditions in the field, with somewhat poorer accuracy achievable in natural streams". Buchanan and Somers (1969) recommend the use of chemical dilution gauging "for those sites where conventional methods cannot be employed owing to shallow depths, extremely high velocities, or excessive turbulence". In New Zealand the main application of the technique has been in measuring small turbulent mountain streams (Keller, 1967) and Water and Soil Division staff have avaluated the technique on alpine streams in the Canterbury high country. During 1977 the technique was applied to calibrating power stations in the Upper Waitaki power scheme.

The dilution methods are an important flow measurement tool, partly because of their potential for high accuracy and also because they can be applied in conditions unsuitable for conventional methods.

WATER–LEVEL RECORDING

As stated earlier, it is customary to measure water–levels (stage height) continuously, then relate discrete measurements of discharge to the stage record to produce some form of continuous discharge record, for example a record of daily mean discharge for a particular station on a river.

From 1900 to 1960 a wide variety of float operated water–level recorders were used in New Zealand. Some of these, Kent and Lea chart recorders, are still in use today. During the 1960's digital 16-track punched tape recorders were introduced and these are now used on all the more important stations throughout New Zealand.

New Zealand experience with punched tape recorders has emphasised the value of having a single specialised instrument service depot. The Water and Soil Instrument Service Centre (WSISC) Christchurch, checks over new instruments and carries out regular maintenance and repair of recorders returned from the field. Initial development problems experienced with digital recorders were more easily overcome because of the technical skills centred at WSISC. Since 1964 digital recorders have been progressively improved by using more satisfactory clocks and, more recently, solid state timers. Motors and other components have also been modified to provide a very reliable instrument.

A survey of stations equipped with digital recorders showed that with monthly inspection visits, less than 4% of the total record is lost because of instrument failure or malfunctions. WSISC provides an instrument repair service to all agencies involved in hydrological work in New Zealand.

At present, water–level recording in New Zealand is carried out using:

(a) float-operated digital punched tape recorders. (e.g. Leupold & Stevens, Fisher & Porter).
These recorders are now very reliable electro-mechanical devices. A major advantage of digitial recorders is that the basic unit of stage measurement is one millimetre and this can be recorded over the full range of the station or river. At many stations in New Zealand the digital recorders are geared with a float driven pulley so that 3 mm of stage is the effective unit of measurement;

(b) float operated chart recorders, for example the Stevens F, Stevens A71, and older Kent and Lea recorders;

(c) pressure - bulb operated circular chart recorders, e.g. Foxboro, Cambridge, Bristol, Kent and Teltherm. These recorders are easy to install, require a minimum of housing and have a flexible tube connecting the recording instrument to a pressure-bulb in the river. Disadvantages are the circular chart and less accuracy. Pressure bulb recorders constitute about one third of all water–level recorders used in New Zealand. Most are used at temporary investigations stations where only a short record is required and at difficult sites where the cost of constructing a recorder float tower has been prohibitive.

At sites where physical conditions permit the construction of a recorder tower at reasonable cost, this method of obtaining a water–level record is still the most precise and satisfactory means available.

In New Zealand towers have been constructed from a great variety of materials including concrete, timber, steel and fibreglass. Towers built before 1960 were commonly reinforced concrete, either square or circular. They are permanent structures and can be a problem should the river change its course or, as is more common, degrade the bed to leave the intake pipes out of water during low flow periods. At some stations degradation of the river bed has been accentuated through shingle extraction.

In recent years the 'standard' tower has been constructed from 1.83 m sections of galvanised, flanged steel tubing, with a diameter of 0.559 m. Special sections are available with inspection hatches and lugs for attaching a cat walk. The cat walk, and angle iron braces fixed to anchor blocks on the river bank give stability to taller towers. A major advantage of these steel section towers is that they can be unbolted and removed to a new site, should the station be closed.

CONCLUSION

The measurement of river flow and recording of water–levels in New Zealand is very similar to accepted practices in North America and Europe, though New Zealand agencies make use of rather less permanent water–level recording towers than is often the case elsewhere. This country has been very successful in using punch tape water–level recorders which produce records suitable for rapid computer processing.

The availability of a central instrument service depot has been a major factor in the successful use of advanced types of water–level recording equipment.

We wish to emphasise that there is no single method of stream flow measurement which should be used exclusively. The real skill in streamflow measurement is to select the best site available and apply the most appropriate measurement technique to the physical conditions prevailing. Other factors such as the purpose of the measurements, the accuracy required, and the duration of the investigation or research will also have some influence on choosing a method of streamflow measurement.

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AUTOMATION IN HYDROLOGY WITH PARTICULAR REFERENCE TO DATA PROCESSING AND CONCEPTUAL CATCHMENT MODELS

R. P. Ibbitt

ABSTRACT

Reasons for collecting hydrological data and for using computers to archive it are given. The range of instrument types used and their different forms of recording data are described. The steps needed to convert the many different forms of field document to a common computer based form are indicated along with the steps needed to ensure data reliability and user access to the data. The basis of one particular use of the data, conceptual modelling, is presented with a review of the developments in New Zealand.

INTRODUCTION

Hydrology does not possess a unifying theory nor are many of its individual problems capable of solution by laboratory experimentation. Most hydrological problems have consequently been solved by making comparisons with what has happened under similar conditions at other places and/or other times. This requires records with which to compare present information and techniques for making reliable comparisons.

For many purposes spatial comparisons are regarded as less reliable than temporal comparisons at the same location owing to the largely unknown and unquantifiable differences between the processes controlling the movement of water in nature. Temporal comparisons are not, however, without their difficulties. Uncertainties arise from the infinite variations that can occur as a result of both before and within event differences.

By recording data from as many sites as possible over as long a period of time as possible it is hoped to achieve two objectives: firstly to obtain enough combinations of different conditions to be able to find events which are sufficiently similar to allow reliable conclusions to be drawn about future behaviour; and secondly to distil out the mechanisms controlling hydrological responses and by accounting for much of the variation between different sets of data to reduce the need for recording vast quantities of data.

World experience has already shown that the amount of data required to meet the first objective has far outstripped human ability to record, prepare, assimilate and use, without some form of automation to organise and condense it. The problem of data collection has been aggravated in meeting the second objective by the need to record more diverse data at increasing frequencies to facilitate the identification of individual facets of catchment response.

Not surprisingly the digital computer has figured prominently in the automation of data processing. Computers also have an important part to play in helping to elicit the mechanisms controlling catchment response. Frequently such mechanisms require repetitive application of simple mathematics to large amounts of data before their reliability over wide ranges of data can be assessed. Obviously before this type of scheme is embarked upon the data must be checked for significant errors. Here again the computer is most useful.

In the first part of this paper a description is given of what has been done to organise, condense and check data so that it can be confidently used for making comparisons. The second part reviews a major use (conceptual catchment modelling) of large amounts of reliable data. Conceptual catchment modelling uses a wide variety of data and has developed into a powerful technique of hydrological analysis and one that can be made to go far towards meeting the second objective.

DATA PROCESSING

Hydrology, being an observational science, requires measurements of hydrological phenomena if there is to be a basis for reliable deduction. Since little can be done with single observations it is usually necessary to record a large number of measured values.

Collection

Early hydrological measurements in New Zealand were made manually using standard scales. One of the first series of regular measurements were those of the level of Lake Rotoiti dating from 1904. Two typical characteristics of the early manual records are that they were of relatively slowly varying phenomena, such as lake levels and that they were made at populated locations and usually had some immediate practical significance unrelated to hydrological work. Lake level observations, for example, were made to ensure that there was adequate water for boats to berth at wharves.

More rapidly varying phenomena, including river levels, required too much manual recording effort to be made on a regular basis. Nevertheless during times of extreme events, flood or drought, readings were sometimes made out of either curiosity or practical concern for the safety of buildings.

The practical use of the data gathered manually stimulated the use of the information for design work and this led on to demands for more and better data. Two factors were of concern at this stage: how to obtain more reliable data, say without breaks because of observer illness; and how to obtain data at remote locations. These requirements led to the introduction of chart recorders, mainly on lakes with significant potential for hydro–power. In the absence of alternative techniques, chart recorders in one form or another served as the chief source of hydrological data from the early 1920s through to the middle 1960s. Their decline was brought about by the need to transfer the chart data to computers for use. The only way this could be done was by manually reading the charts, an initially skilled, time consuming and inaccurate process. Development of x–y digitisers considerably reduced the demand on human involvement but did not eliminate it. The situation deteriorated further when increasing demands for higher resolution on charts resulted in instruments that either produced more charts, and/or incorporated devices for 'folding' charts that made reading more error prone.

During the 1960s many chart recorders were replaced with 16–track paper tape digital recorders. These instruments are able to punch data on to paper tape that can then be read by machine and converted into output directly readable by computers. The intermediate step is necessary because 'high speed' input data to a computer must meet exacting standards for which 16–track paper tape was not designed.

The problem encountered with the commercially available digital recorders is that their criterion for punching a phenomenon value is the elapse of a fixed time from the previous punch. For the more common hydrological variables this form of recording is not completely satisfactory since it produces a glut of data when little is required (during times of inactivity), and a dearth of data when high time resolution is required (during a flood). The usual compromise is to record at the highest frequency considered to be necessary to capture the required information. According to Chandler and Patterson (1970) this can result in punching up to 20 times more numbers than are necessary to adequately define the changes in phenomenon value.

During the late 1960's and early 1970's the Ministry of Works and Development modified the punching criterion of one of the standard makes of 16–track paper tape punch recorder to reduce the over–recording. The modification used for recording water levels, punches data according to the rate of change that it senses. The greater the rate of change the more frequently the instrument punches, up to a predetermined maximum frequency. If there is no activity the instrument behaves like the standard recorder except that the punch interval is much longer.

For rainfall an equivalent instrument was developed. Punching is initiated when a fixed amount of rainfall (the trigger value) has been collected. The quantity punched is the time of the end of the small fixed–length interval in which the amount of rainfall reached its trigger value. In all cases the sensor feeding data to this instrument has been a tipping bucket raingauge although in theory any sensor of a similar kind could be attached.

For details of both types of event recorder see Chandler and Patterson (1970). Patterson (1973) gives more information on the water–level recorder.

Combining the principles used by both types of recorder Patterson (1971) developed a multi-channel recorder capable of recording data from several different sensors. At the end of each small fixed–length time interval, the data channels are scanned to see if any has reached its trigger value. If a trigger value has been reached the machine punches a channel identification and the elapsed time from the previous time check.

The main difference between these three types of event recorder and commercially available equipment is that a limited amount of 'intelligence' has been introduced into the event recorders to screen the data for redundant values and eliminate them before recording.

The development of new instruments is seldom free from problems and the event recorders are no exception. Of those in service the event rainfall recorder has undoubtedly proved the most successful and is now in use throughout New Zealand. The event water level recorder has had a more chequered career, the major difficulty arising from its own efficiency in eliminating redundant data. As long as there are no instrument malfunctions and the field checks are conducted correctly, records of high quality are returned. However, whenever anything goes wrong the absence of redundant data often makes it impossible to salvage any useful information. For these reasons event water level recorders are best used for experimental purposes where regular checks can be run on instrument behaviour and where high quality data are required. The setup at the Moutere Farm experimental station (Ministry of Works and Development (1971)) is typical of the arrangement in which event water level recorders give best results.

Of the three event recorders, the multi–channel recorder has seen least service. While its design and development were able to take advantage of the problems encountered with the other two event recorders, there have been fewer applications for its capabilities. Some data from one of these instruments have been processed but are insufficient to judge its success as a routine field recorder.

As a result of the experiences gained primarily with the event water level recorder, it is now recognised that where temporal resolution of the data is not of paramount importance, the standard commercial recorders give easier processing and more reliable performance. With modern data processing techniques redundant data can easily be removed during processing, greatly reducing the need for economy of recording at the station.

Preparation for Archiving

The purpose of data adjustments is to convert the multitude of forms in which data are recorded in the field into a common form incorporating data corrections detected as a result of manual checks on instrument performance. The amount of data being collected on a national scale can be stored, organised and retrieved only by computers. In New Zealand this is done by the Water and Soil Division of the Ministry of Works and Development using the TIDEDA data processing system (Ibbitt (1972a), Thompson and Wrigley (1974, 1976).

Early in the development of a data processing system suitable for hydrological data, a decision was made to adopt a form of data storage that would retain as far as practical, all the information in the original field document. In this way artificial restrictions on data use are not imposed by the processing system, but at the cost of providing retrieval facilities that can put any desired interpretation on the data. As a consequence of this decision an assumption was necessary about how the data varies between recorded values. The assumption of linear variation is made so that there is consistency between the processing system and the basis for determining the punch interval of digital recorders and the points on chart traces to be digitised.

Implicit in the decision to store data so as to be able to reconstruct the field document is the decision not to rate water levels before filing. To obtain flow data the ratings are stored as physically separate entities which can be retrieved and applied when required. In this way rating data can be retrospectively changed without having to reprocess the water level data. When retrieving data the system has facilities for altering its form to suit user requirements (e.g., daily mean values may be retrieved).

The large quantity of data collected for storage has already been referred to. To avoid incurring unnecessary overheads during processing all data are subject to a compression process which removes items that can be defined adequately by linear interpolation, as discussed in Ibbitt (1975a). While data compression works well with river level data, it is not as successful with oscillating water levels recorded at high frequency. Such data arise from lake seiching and tidal effects. Thompson and Ibbitt (1978) describe how data subject to wind seiching are treated while Gilbert (1978) describes a technique for overcoming the more difficult problem of barometric seiches.

For data affected by tidal influences the situation is somewhat different since the amplitude of the oscillations is generally of comparable magnitude to those caused by other natural phenomena. Where the maxima and minima of the tidal variations are regarded as important the data are being left as initially processed but where tidal influences obscure important information on long term trends, the tidal cycle is removed. The latter cases may arise for artesian groundwater data.

Editing

During the collection and archiving of hydrological data a variety of errors can be introduced by data sensor, instrument malfunctions and human errors. The purpose of editing data is to remove these errors and to provide ancillary information such as ratings and comments needed to explain unusual aspects of the data. Data errors fall into three main categories.

1. Isolated individual values are grossly in error e.g. a mispunch by a digital recorder.
2. Small batches of data consisting of all values from one or two field documents are in error (e.g., a group of charts may have been processed using incorrect instrument constants).
3. Long periods of data are subject to small, perhaps intermittent, but cumulatively significant errors (e.g., an unsuspected blockage of the bottom hole of a recorder intake pipe may affect the lower parts of hydrograph recessions leading to an inflated estimate of the available water resource).

The first type of error is the most easily detected and except for extreme value analysis is of little practical importance. Such errors have received a disproportionate amount of editing effort largely because they tend to discredit, in the eyes of the inexperienced, the value of an otherwise good record. Errors of the second and third type have more serious practical consequences. Fortunately the second type of error can be easily detected and corrected. However, the third form of error is not easy to detect and if found it may not be possible to correct it.

Besides correcting data for errors the other main aspect of editing work is to add information not readily available from automatic data recorders. Such information can range from rating data for use in converting water levels into flows, through sediment ratings and cross–section data, to comments giving details of data irregularities.

The techniques for dealing with all three error types and for adding and checking additional information are described in detail in Thompson and Wrigley (1974) and Ibbitt (1977).

Access

Access to hydrological data has always been a thorny issue. Seldom has it been possible to meet the conflicting needs of:

1. Ready access to information in a form suitable for direct and perhaps sophisticated calculation;
2. Wide availability to all potential users;
3. Minimum time delays between collection and dissemination;
4. Freedom from restrictions on data form. For example if data are collected at 15 minute intervals but only monthly values are released for general use many potential users will reject the data as inadequate.
5. Amendments should become automatically available to the user.

Ibbitt (1975b) discusses in detail the system in use in New Zealand. The approach has removed the overriding weight historically given to item two and goes a long way to meeting the other four items.

The main difficulties with the present system are a lack of awareness of the system on the part of some potential users and a lack of suitable computer facilities on the part of others. Many of these problems are being overcome with the passage of time, and the increasing use of magnetic data tapes by consulting engineers is a good omen for the future.

The improved flexibility of the present system is not without its costs. The former dominance of annual yearbooks with their crippling restrictions on data form, tended to obscure deficiencies and changes in data accuracy and reliability. Abolition of this restriction has placed a far greater onus on the users to ensure that data are suitable for their purpose. To help meet this need the added comments in the data files give a 'picture' of reliability and adequacy (Fig. 1). Any queries about the data should in the first instance be referred to the local recording authority who hold copies of the *Flow Reports* for fully edited data. A *Flow Report* details the editing work done on the data and includes many of the frequently requested statistics.

DATA USE: CONCEPTUAL MODELS OF CATCHMENT HYDROLOGY

The availability of data in computer compatible form has led to a wide range of applications of the computer to the solution of hydrological problems. Initially the computer's speed was exploited to accelerate analysis formerly done by hand. Certain techniques, particularly those involving subjective decisions and graphical analysis, proved unreliable, if not dangerous, when computerised and released to inexperienced users. Dissatisfaction with the computerisation of some manual methods coupled with a growing realisation that the computer permitted development and application of new but previously impractical ideas led to new computer based techniques. One of the more elaborate and promising of these is the field of conceptual modelling of catchment hydrology.

FIG.—1 Comments filed with the data.

13 90604 0 0 710522 240000Comment
Initial comment for Hokitika River at Colliers Creek
Site number 90604 on river number 906000
The site is situated 37 kms from the mouth at grid reference S57:549228
Drains 352 KM2 and control is natural
Additional information: sediment concentration also measured at this site
The local recording authority is the Christchurch hydrological survey
13 90604 0 0 710523 240000Comment
F & P recorder installed on day 710524
13 90604 0 0 720110 240000Comment
Synthetic record from 720111 0 to 720718 121500
The recorder adjustment screws were loose during this period therefore all records were faulty
The synthetic record was derived by correlating this site with the Butchers Creek flow site (90605), grid reference S58:630587, and two rainfall sites: Rapid Creek (automatic) (310010) and Kawhitirangi 2 (Met. F21903)
13 90604 0 0 730112 240000Comment
Synthetic record from 730113 0 to 730327 101500 Records or file between 730113 0 and 730205 211500 were faulty
Clock stopped 730205 211500.
Synthetic record was derived by correlating this site with the Butchers Creek flow site (90605), grid reference S58:630587, and Hokitika Catchment mean rainfall
13 90604 0 0 751015 114500Comment
F & P recorder removed day 751015
L & S 16 track punched paper tape recorder installed day 751015
13 90604 0 0 760619 100000Comment
Recorder out of action between 760619 at 100000 to 760621 at 143000 owing to a blocked static pipe. Finally cleared and flushed on 760621.
Record derived by correlating with the rainfall from the Rapid Creek (automatic) (310010)
13 90604 0 0 760805 131500Comment
Clock stopped 760805 at 131500.
Synthetic record covering period 760805 at 131500 to 760916 at 120000 was derived by correlating with the Westland Catchment Board Foxboro records from their Hokitika at S.H.B. site. Reliability: good

Note:

1. new comments begin with the word COMMENT at the extreme right hand end of the line;
2. of the 6 numbers that precede the word COMMENT, the second is the station number, the fifth is the date of the comment in the form year/month/day and the sixth is the time of the day in the form hour/minute/second.

The basis of conceptual modelling is to treat a catchment as being made up of a series of storages. Movement of water between storages is related to the laws of physics. The degree of rigour with which physical laws can be incorporated depends on four factors: accurate knowledge of which laws govern the inter–storage movement of moisture on a catchment wide basis; quality of the data for checking the reliability of the assumed mechanisms (i.e., storage plus transfer mechanism); significance of each component in the overall response of the model; availability of facilities for implementing calculations.

While the qualitative significance of moisture transfer from precipitation to runoff has long been manifested in the concept of the hydrological cycle, it is no coincidence that the quantification of this concept had to await the availability of digital computers. The first major development was the Stanford Watershed Model developed by Linsley and Crawford in the late 1950's. The potential of this development was quickly grasped and seemed to offer the promise of a hydrological panacea.

One of the earliest difficulties encountered by conceptual modellers, and one which remains, is the quantitative estimation of the parameters controlling the operation of individual storage/transfer mechanisms. For example, in routing flow down a river channel some measure of channel roughness is required. Meaningful direct measurements of roughness are not possible for real channels although estimates of average values can be deduced.

Two approaches have been made to overcome this problem. Firstly experiments have been carried out to allow accurate deduction of parameter values. Such experiments usually try to isolate one component of the hydrological cycle, measure input to the component, output from it and deduce storage properties from the results. The cost of such experiments necessarily limits their size and most have been carried out on small plots believed to be representative of average catchment conditions. Greatest success has been achieved where effective isolation of a component was possible, e.g. laboratory experiments on channel routing and field experiments on interception (Blake (1972)).

The alternative to experimental estimation has been to employ iterative numerical techniques. By assuming an initial value for each parameter, rainfall can be 'routed' through a model to form runoff. Knowledge of the actual rainfall, runoff, and potential evaporation, then permit the calculated and measured runoff to be compared and systematic discrepancies can be interpreted as errors in the parameter estimates. Parameter values are then adjusted, the routing process is repeated, and new assessments of the results made.

Developments in New Zealand

If there are more than a few parameters the estimation of their values by repetitive calculation becomes tedious. Computerised numerical techniques are available to take the labour out of the estimation of parameters and were used by Boughton (1968), Murray (1970), Taylor (1972) and Ibbitt (1976). Boughton's model was a development of earlier work carried out in Australia (Boughton, 1965; 1966) and used a steepest descent method to fit a model of daily values to five small catchments in New Zealand. Murray took an elaborated form of Boughton's model and applied it to daily values from a small catchment in North Wales, while Taylor, using data from six representative basins, fixed the values of the less sensitive parameters to reduce fitting problems. One of the innovations in Murray's work was the use of a split record test to assess model reliability. In a split record test the model is fitted to one period of data and is then used to 'hindcast' a period of data not used in the fitting. In this way a model's ability to extrapolate outside the conditions imposed on it by the fitting period can be assessed. Ibbitt (1976) aimed at producing estimated runoff to a predetermined level of achievement. Although the objective was not met an analysis of the differences between measured and calculated values demonstrated that this was caused by data limitations rather than model short-comings. Similar problems seem to have severely restricted the application by Hutchinson and Simmers (1971) of the Boughton model to the Upper Taieri basin, again using daily values.

Use of sub–daily information was introduced by Wood and Sutherland (1970) and Wood (1973) who chose to modify a version of the Stanford Watershed Model and to fit data manually at quarter hourly and hourly intervals. The emphasis in their work was to replace much of the empiricism of the infiltration procedure of the Stanford Watershed Model with a more realistic infiltration component based on a numerical solution to the one dimensional Richards equation for flow in an unsaturated porous medium. The change reduced the number of parameters, always a worthwhile objective in catchment modelling, while improving the model's physical basis. The use of time intervals of less than one day brought in complications because it placed more stringent requirements on all the model components to forecast more accurately the temporal changes in moisture storage.

Of the models discussed all are classified as lumped models since the components for each identifiable part of the hydrological cycle have been assumed to be spatially representative of catchment conditions. Furthermore the characteristics of each component have generally been deduced on intuitively reasonable grounds and very little component testing has been carried out owing, in a large part, to the difficulties of getting reliable data.

With the exception of the general component model of interception described by Ibbitt (1972b) other component models have arisen as a result of special investigations where particular components of the hydrological cycle are dominant. One of the most notable in this category is the series of models developed by Donaldson (1974) and Donaldson and Campbell (1977) for studying the behaviour of the Lower Hutt groundwater system. The final model provides a finite element solution of the two dimensional differential equations describing movement of water in both the unconfined and confined parts of the aquifer.

Wooding (1966) describes the development of a flood routing model which accounts for infiltration but concentrates on the hydraulic aspects of routing precipitation excess to the catchment outlet. The deceptive ease with which surface water can be modelled and its relative importance in the urban situation has resulted in this type of model receiving far more emphasis overseas than in the predominantly rural environment of New Zealand.

In contrast New Zealand has put a great deal more effort into trying to solve the problem of rural catchment behaviour where dominance by one particular component is not so great. One of the main hopes for the application of conceptual catchment modelling has been the quantitative assessment of changes in land use. Until the component parts of full models can be tested, or at least their behaviour within the model verified indirectly as in accordance with the few observations that are available, the goal of a model capable of giving reliable quantitative assessments of the effects of land use changes will be elusive.

Beable (1976) and Jowett and Thompson (1977) describe the use of distributed catchment models for modelling large catchments. In these models the catchment is broken down into a number of sub–catchments. For each sub–catchment a rainfall excess is derived from the storm rainfall using a loss rate concept, and is then subjected to a mechanism to simulate its travel to the sub–catchment outlet. For this Jowett uses unitgraphs (either derived from available records or estimated using Snyder's (1938) method). The greater degree of subdivision required by Beable, coupled with fewer gauging stations, resulted in the rainfall excess being directly translated to surface runoff which was then routed to the sub–catchment outlet. The routing coefficients depend on the topography and channel slope. Outflows from sub–catchments are combined and routed downstream as kinematic waves (Jowett) or using a Muskingum type of equation (Beable).

Although similar in concept the models of Jowett and Beable have considerably different purposes. Jowett is primarily concerned with obtaining flood estimates for design purposes and assumes no major changes in land use. On the other hand Beable's work is primarily aimed at allowing qualitative assessments of land use changes on flood discharges. Although Beable's model allows 'before' and 'after' comparisons, its present state of development restricts predictions to being better than or worse than some alternative. Collection of 'before' and 'after' data should permit refinement of the comparisons and could lead to quantifiable assessments.

Besides examining changes in flood discharges, Beable's work can be combined with sediment ratings to give qualitative information about land use changes on the sediment yield aspect of water quality. This extension does, however, make the unlikely assumption that a change in land use does not affect the rate of erosion, only the amount of water available to move the results of the erosion.

A discussion of conceptual catchment modelling in New Zealand would not be complete without mention of work done on water quality modelling of rivers. This type of modelling has been particularly problem orientated, with work by McBride (1976) on the Mataura, Rutherford (1974) on the Waikato and Rutherford and O'Sullivan (1974) on the Tarawera being the most notable. In these works biological oxygen demand and dissolved oxygen concentration variations along the stream channel are calculated using the governing differential equations. While this work can be looked at as a field on its own it is capable, if required, of incorporation into some of the more general catchment models previously discussed.

The Future

The early emphasis on catchment modelling was to produce estimates for planning purposes. Protection of the environment as well as property and lives, is moving the emphasis towards operational forecasting models. Improvements in data collection foreseeable within the next few years will hasten this change. No doubt new and more specific models will be required in man's incessant search for better forecasts of what is about to befall him. In the quest for these better models there will be the temptation, particularly when using automatic fitting techniques, to manipulate more components rather than comprehend the interaction of fewer. This temptation must be resisted if the foundation of hydrological modelling in New Zealand is not to be eroded in the interests of expediency.

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UPSTREAM GENERATION OF STORM RUNOFF

A. J. Pearce and A. I. McKerchar

ABSTRACT

Storm runoff volumes and net storm rainfall amounts are used to calculate runoff-contributing areas for a range of storms on 17 small catchments in nine regions distributed through New Zealand. Relationships between depth of storm runoff and the proportion of each catchment contributing to storm runoff indicate that the relative importance of different runoff-producing mechanisms varies in time and in space. In all catchments studied, storm runoff during small events (return period 1 day – 100 days) can be explained by saturation overland flow on small proportions of the catchment. The mechanism generating runoff in larger events varies among catchments in relation to topography, geology, climate, and the hydrogeologic characteristics of the underlying soils and rocks.

For three catchments with low infiltration rates, moderate–sized events are generated by partial–area (Horton) overland flow, and the largest events are generated by widespread Horton overland flow. On one catchment underlain by thick Recent and Pleistocene tephra, a sharp, initial hydrograph peak is derived from saturation overland flow, but the bulk of storm runoff in large winter events does not reach the channel until several days after rainfall. This broad secondary peak with slow rise and slow recession is almost certainly generated by subsurface flow. For all other catchments, regardless of vegetation type or terrain, the bulk of storm runoff in larger events is generated by rapid subsurface flow from between 40% and 90% of the catchment area.

Storm runoff yields from the catchments studied are large in comparison with yields reported from elsewhere; the explanation lies in the combination of steep slopes, thin soils of high saturated hydraulic conductivity, and incised stream channels which are typical of New Zealand hill and mountainlands. Theoretical support for the high yields of rapid subsurface flow is found in the mathematical models of Freeze (1972, 1974), which couple simulations of three–dimensional, transient, saturated–unsaturated subsurface flow, with one–dimensional, unsteady, gradually–varied channel flow.

INTRODUCTION

During the last decade, substantial theoretical and empirical advances have been made by North American and European workers in explaining the production of storm runoff. These advances stemmed in part from a dissatisfaction with the adequacy of the concepts of Horton (1933, 1939, 1945) in accounting for the production of storm runoff in humid areas, particularly under forest cover. The development of the concepts of partial area (Betson, 1964) and variable source area (U.S. Forest Service, 1961) production of storm runoff has been reviewed by a number of writers, e.g., Dunne (1976) and Freeze (1974), who also reviewed the theoretical and mathematical model background to studies of storm runoff production.

The new concepts of runoff production are now beginning to be used in catchment modelling and land management studies (e.g., Engman and Rogowski 1974, Dunne *et al.* 1975). These developments in both theory and practice have, however, received little recognition in the New Zealand literature, and it would appear that their implications have not been fully recognised by managers of water resources and land. Much land management is still predicated on an assumption of spatially uniform runoff production via the Horton model.

One notable exception is the attempt by Hayward (1976) to explain storm runoff from a Canterbury mountain catchment using a model based on variable source area contributions of subsurface stormflow.

In this paper we examine some evidence bearing on the relative importance of the various mechanisms of storm runoff production in upstream areas for a number of small catchments in New Zealand. The range of soil, geologic, topographic, and vegetation conditions does not fully cover the wide range of hydrologic conditions found in New Zealand. Nevertheless, we believe that this study shows that various runoff production mechanisms are probably differentially important among regions, and that care is needed in extrapolating results from elsewhere to some areas of New Zealand, especially those where the terrain is steep, streams are incised, and the soils are shallow but highly permeable.

CONCEPTS AND TERMINOLOGY

In general we have followed the use of terms as defined by Freeze (1974) and Dunne (1976). The history of the development of the concepts of Horton overland flow, the partial area hypothesis, and the variable source area concept is clearly summarised by Freeze (1974) and does not need repetition here.

Storm runoff can be produced by several mechanisms which deliver runoff to a stream via different pathways. The relative importance of these pathways depends on climate, geology, topography, soil properties, and vegetation cover. Three basic mechanisms which appear to satisfactorily explain storm runoff production in all combinations of the above controlling factors have been verified by field observation at various locations:

(1) Horton overland flow is produced when rainfall rates exceed the infiltration rate of the soil and the upper parts of the soil profile become saturated. The excess of rainfall rate above the saturated hydraulic conductivity of the soil becomes available for surface detention and flow over the ground surface. For a given rainfall rate on a given catchment, only some soils will have infiltration rates less than the rainfall rate. These areas are called partial areas, and their distribution is controlled by the distribution of soil types.

(2) Saturation overland flow is generated by rainfall on saturated areas near stream channels and in valley floors. Rising water tables (fed partly by rainfall that has entered the soil upslope of the runoff source area) reach the ground surface soon after rainfall begins, and further rainfall on the saturated zone generates flow over the surface. The saturated area expands and contracts during each storm and seasonally. The principal controls on the distribution of saturated areas are the topographic and hydrogeologic characteristics of the hillslopes in a catchment.

(3) Subsurface flow is generated by rapid infiltration of rain and the consequent increase in hydraulic conductivity of the soil as moisture content rises. The infiltrated water moves laterally through permeable horizons or in perched saturated zones directly into the stream network which expands headward and laterally as rainfall progresses and valley–floor saturated zones develop. The stream network expands and contracts seasonally as well as during individual storms.

Research during the last decade in North America and Europe has indicated that Horton overland flow is probably only important in arid and semi–arid climates, on land used for arable farming, and where infiltration rates are low for whatever reason. The generation of storm runoff in humid areas is still the subject of controversy, especially on forested lands. Dunne (1976) has attempted to clarify this controversy by proposing a gradation of importance of the two variable–source mechanisms. In areas with thin soils, gentle topography with concave lower slopes and wide valley bottoms, Dunne suggests that saturation overland flow dominates the storm hydrograph. In areas with steep straight slopes, incised channels, narrow valley bottoms, and deep very permeable soils, subsurface flow dominates the storm hydrograph.

METHODOLOGY

The storm hydrographs studied in this paper are from small catchments where the size and shape of the hydrograph are controlled by the input mechanisms and are relatively unaffected by channel storage. The catchments used in this study range in size from 2 ha to 52 ha; even in the largest of the study catchments, channel storage should play only a minor role in determining the shape of the hydrograph.

For consistency in computing stormflow we have used the method first proposed by Hewlett and Hibbert (1967), where a separation line sloping at 0.0055 1/sec/ha/h (0.05 cfs/sq.mi./h) is projected from the initial rise until it intersects the receding limb of the hydrograph (see Fig. 1 for hydrograph parameters). This method, although somewhat arbitrary, separates quickflow (stormflow) and delayed flow but makes no assumptions as to the mechanism of quickflow production. It is now widely used in both Britain and North America because it allows a ready comparison of hydrograph parameters that have been computed on a uniform basis for catchments in different hydrologic regions (e.g., Corbett *et al.* 1975; Gregory, 1974; Harr *et al.* 1975; Harr, 1977; Walling, 1974).

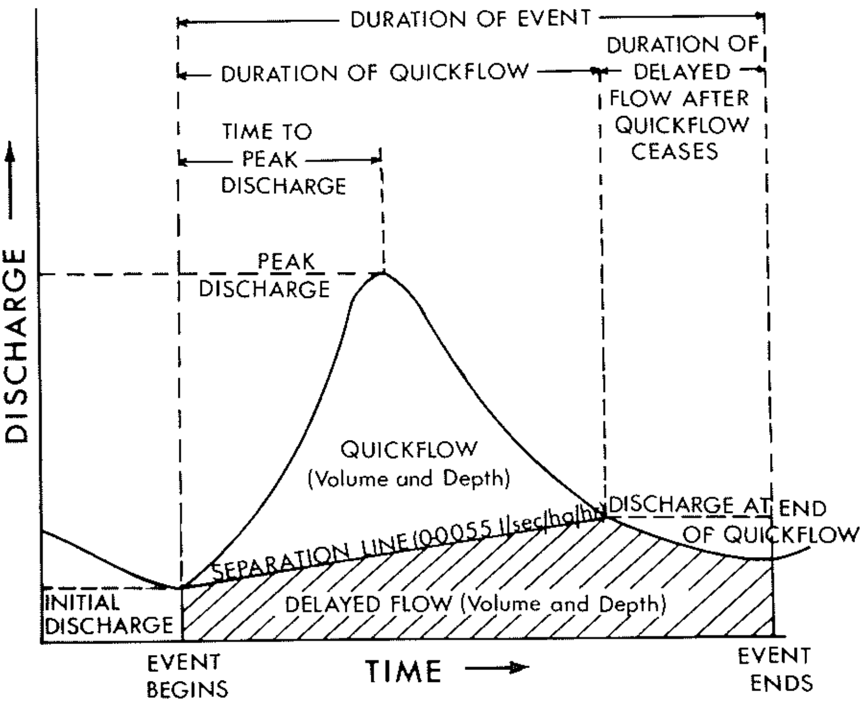

FIG.—1 Hydrograph separation parameters, Hewlett and Hibbert (1967) separation method.

Our treatment of the data from each catchment has depended on a number of considerations other than catchment size. For catchments monitored by the

Water and Soil Division, M.W.D., annual rainfall and runoff records for the whole period of record were examined and a "typical" year selected. The complete hydrograph for that year was plotted and quickflow amounts and total runoff were calculated using the M.W.D. digitiser/calculator system. The total runoff was then compared with the TIDEDA record for that year to eliminate gross errors. (M.W.D. filed data system, see Thompson and Wrigley (1974)). Because of plot–scaling problems, and the precision of area measurements, the two runoff totals differ in some instances by as much as 10% but typically differ by 10–20 mm or a few percent of the runoff total. We assume that these errors are distributed across all the events in each year of records, thus quickflow yields for individual events should not be in error by more than a few millimetres of runoff even in the largest events. Runoff amounts digitised from the plotted hydrograph were used in all calculations, but TIDEDA figures for total runoff are given for comparison in each section as well as in Table 1.

Hydrograph records from catchments monitored by the Forest Research Institute are routinely separated, thus hydrograph parameters for periods ranging from 14 months to 7 years were available for these catchments. The varying availability of other data such as rainfall amount and intensity, infiltration rate, and soil permeability has enforced other limitations on our interpretation of the runoff yields which are noted in more detail in the discussion of individual areas.

Stormflow response for each catchment was investigated by expressing quickflow yield as a percentage of rainfall, either annually or for individual events. This approach is not new but derives from analyses by Betson (1964), Hewlett and Hibbert (1967), Ragan (1968) and, more recently, Harr (1977). The proportion of rainfall yielded in quickflow for a single event can be viewed as an estimate of the proportion of the catchment contributing to stormflow (U.S. Forest Service, 1971; Hewlett, 1974; Harr, 1977). We have chosen to use net storm rainfall for individual events (rather than gross rainfall as in Harr, 1977) because conceptually this would appear to provide a more realistic estimate of the contributing area. Either approach provides a minimum or near-minimum estimate since the method assumes 100%delivery of the rainfall on the area so identified. Net storm rainfall amounts were estimated using published regression relationships between gross and net rainfall for the appropriate vegetation types. Data for most of the vegetation types show highly significant linear relationships between net and gross rainfall for storm period data (Blake, 1972, 1973; Rowe, 1975), which although obscuring the nature of the interception process (especially in smaller events) provide a reasonable estimate of net storm rainfall in larger events.

Table 1 lists the catchments studied and some relevant physical details; catchment locations are shown in Fig. 2. Runoff production is examined for each catchment in the order in which they appear in Table 1.

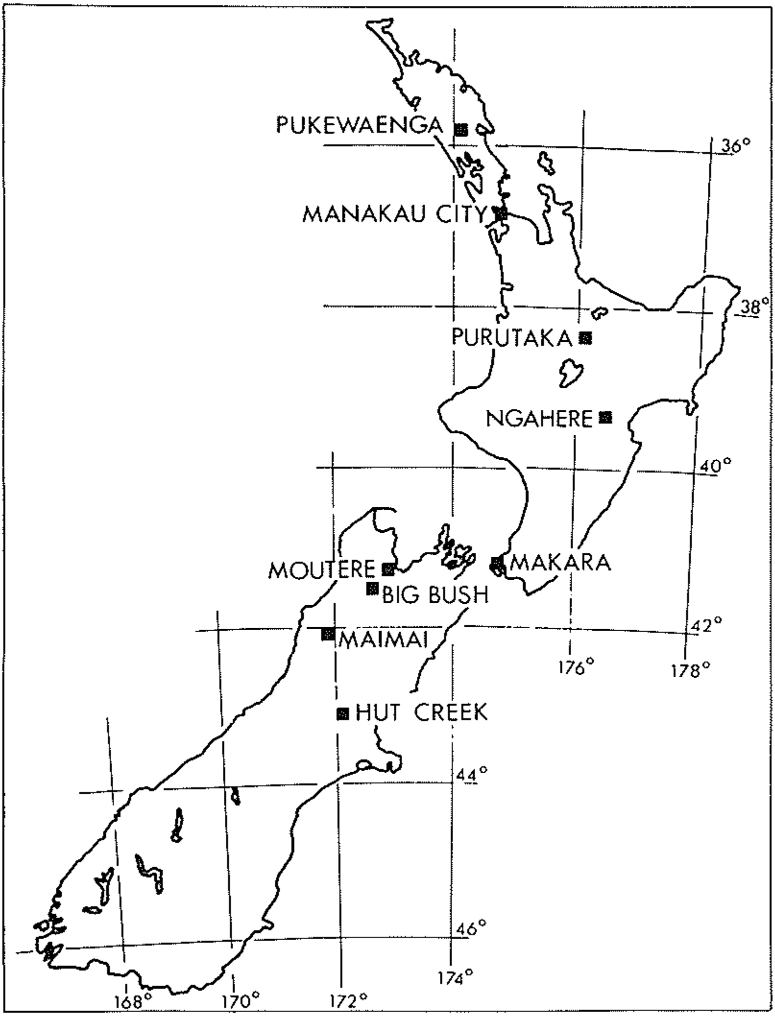

FIG.–2 Location of catchments.

STORM RUNOFF PRODUCTION

Pasture cover on rolling–steep hill country, Pukewaenga catchment

Pukewaenga catchment is one of three hill–country catchments in the Puketurua experimental basin 20 km north–west of Whangarei. The 38.9 ha catchment was monitored under scrub cover from 1966 to 1972, when it was converted to rye/clover pasture. Details of the catchments and the experimental programme are available (M.W.D. 1968a, 1970a, 1971a, 1973a) and Schouten (1976) summarises the results of the experimental treatments. Quickflow yield in 34 events during 1974 ranged from 0.3 mm to 70 mm, and total quickflow was 266 mm. A typical storm hydrograph is shown in Fig. 3. Total runoff digitised from the plotted hydrograph was 611 mm (TIDEDA total runoff is 548 mm). Quickflow was 43.5% of total runoff, and 21% of gross rainfall. Interception under pasture

TABLE—I Physical characteristics of catchments studied

Hydrological Location	General	Catchment (ha)	Area (mm)	Annual Rainfall (mm)	Annual Runoff
Hokianga	Whangarei hill country	Pukewaenga (Puketurua)	38.9	1380	570
Pukekohe	South Auckland	Manukau	30.1	1150	450
Taupo Rhyolite	Central Volcanic Plateau	Purutaka (Purukohukohu)	22.5	1460	360
Kaweka	Kaweka Range	Ngahere	52.0	2690	1760
Makara	Wellington hill country	Makara 11	7.4	1190	330
Moutere Hills	Moutere Hills	Moutere 5	7.0	1120	270
Inland Motueka	Hope Saddle	Big Bush 1	8.6	1800	)
"	"	Big Bush 2	4.8	"	) average
"	"	Big Bush 3	7.8	"	) 800
"	"	Big Bush 4	20.2	"	)
Buller	Central Grey-Inangahua Depression	Maimai 5	2.3	2600	)
"	"	Maimai 6	1.6	"	)
"	"	Maimai 8	3.8	"	) average
"	"	Maimai 13	4.3	"	) 1550
"	"	Maimai 14	4.6	"	)
"	"	Maimai 15	2.6	"	)
Canterbury Foothills	Craigieburn Range	Hut Creek	22.1	1500	750

Geology	Soil Type	Vegetation	Topography	Length of Record	Record Period Used
Sandstone and claystone chaos-breccia (Cretaceous and Tertiary	Northern yellow-brown earths (silt and clay loams)	Pasture (grazed)	Rolling to steep hill country	11 years (TIDEDA)	1974 (Rainfall 1278 mm runoff 548 mm)
Alluvium overlying calcaerous sand-stone (Tertiary)	Manurewa silt loam + hill soils variants (Northern YBE)	"	Gentle, rolling terrain	7 years (TIDEDA)	1975 (Rainfall 1314 mm runoff 481 mm)
Pumice breccias and tuffs with some ignimbrite (Pleistocene and Recent)	Oruanui silty sands + hill soils (yellow-brown pumice soils)	"	Rolling to steep hill country	8 years (TIDEDA)	1975 (Rainfall 1543 mm runoff 384 mm)
Indurated sand-stone and shale (Mesozoic), some thin overlying pumice (Recent).	Steepland soils related to central yellow-brown earth, some pumice soils.	Red beech/ mountain beech forest (virgin)	Steep hill and mountain lands of eastern Kaweka Range	9 years (TIDEDA)	1970 (Rainfall 2630 mm runoff 1661 mm)
Indurated sand-stone and siltstone (Mesozoic)	Makara steepland soils from central yellow-brown earths	Poor pasture and scrub — lightly grazed	Steep hill country	8 years (TIDEDA)	1970 (Rainfall 1215 mm runoff 303 mm)
Weathered early Pleistocene gravels.	Rosedale Hill Soils (sand silt and silt loams) — Central YBE	Pasture (grazed)	Moderate-steep dissected hill country	15 years (TIDEDA)	1975 (Rainfall 1117 mm runoff 272 mm)
"	Hope Hill Soils (stony, slightly pod-solised yellow-brown earths)	Beech-podocarp-hard-wood forest (virgin)	Moderately steep dissected hill country	15 months	All
"					
"					
"					
"	Blackball Hill Soils (stony, podsolised yellow-brown earths)	"	Steep, dissected hill country	27 months	All
"					
"					
"					
"					
Indurated sand-stone and shale (Mesozoic), thin coal measures over-lying (Tertiary).	Bealey Hill Soils (High country yel-low-brown earths)	Induced pas-ture, scrub, mountain beech, conifer mixture	Moderate foothill slopes of Craigieburn Range	7 years	All

cover averages 20% (Blake, 1973), thus quickflow was 26% of net rainfall.

Infiltration rates for six sites in this catchment under sprinkler irrigation ranged from 4 mm/h to 108 mm/h with a grand mean of 42 mm/h (Blake, 1973). Artificial rainfall rates were 53 mm/h to 149 mm/h. Rainfall intensity–duration–frequency data for Glenbervie Forest (20 km distant) from Robertson (1963) are shown in Table 2. Annual maxima for 1969–74 for ten–minute durations all exceed 60 mm/h, those for 30–minute durations exceed 50 mm/h in five out of six years, and in four out of six years 1–hour annual maxima exceed 40 mm/h. It seems clear that rainfall rates for durations up to one hour frequently exceed infiltration rates, thus storm runoff could be generated by Horton overland flow.

TABLE—2 Rainfall intensity–duration–frequency data for various stations, after Robertson (1963)

Station :	*Glenbervie*		*Mechanics Bay*		*Whakarewarewa*		*Taupo*		*Kelburn*		*Greymouth*	
Return period:	*2y*	*50y*	*2y*	*50y*	*2y*	*50y*	*2y*	*50y*	*2y*	*50y*	*2y*	*50y*
Duration	mm/h											
10 min	84	180	75	151	63	125	52	98	41	81	67	142
30 min	51	121	38	82	37	75	29	66	24	48	40	82
1 h	38	?	23	53	26	55	31	47	17	31	27	62
2 h	25	52	15	38	17	36	15	35	12	20	18	35

TABLE—3 Quickflow yield as a proportion of net rainfall, Pukewaenga

Depth of quick-flow (mm)	*No. of events*	*Mean proportion of net rainfall yielded in quickflow (%)*	*Standard error of Mean*
0.3 – 0.99	6	13.3	±2.9
1.0 – 4.99	17	29.7	±3.9
5.0 – 14.99	8	42.9	±5.4
> 15.0	3	60.4	±6.5

Quickflow yields for the 34 events were calculated as a percentage of net storm rainfall for each event (Fig. 4), and averaged for four classes of event size (Table 3). Events yielding less than 1 mm of quickflow produce an average of 13% of net rainfall, but in larger events between 30% and 70% of net rainfall is yielded in quickflow. Schouten (1976) indicates that about 10% of the Puketurua catchment is swampy at all times, thus runoff in events yielding less than 1 mm of quickflow can be explained by saturation overland flow. The combination of limited infiltration rates and high rainfall intensities noted above suggests that storm runoff in larger events is generated by Horton overland flow from various proportions of the catchment, ranging up to 70% of the catchment in the largest storms. The storms yielding less than 1 mm have return periods in the order of weeks (34 events of this size or larger in 1 year). Return periods of the larger events in 1974 are presumably in the order of months to a few years.

Theoretical support for the transition from saturation overland flow dominating the storm hydrograph in small events to Horton overland flow dominating larger events comes from the simulation models of Freeze (1974, Fig. 18, Case C). On poorly permeable soils when rainfall rates are less than infiltration rates, storm runoff is produced mainly by saturation overland flow. In events where rainfall rate exceeds infiltration rates, the simulations show that Horton overland flow dominates the storm hydrograph.

Pasture cover on gentle terrain, Manukau catchment

The 30.1 ha Manukau experimental catchment lies 20 km south–east of the centre of the Auckland urban area on gently rolling terrain underlain by Tertiary sandstones covered by silt loam soils (M.W.D. 1971b). The soils are 50–60 cm thick; typical profiles have 20 cm of silt loam overlying 30–40 cm of silty clay and clay loam which in turn overlies weathered sandstone (Orbell, 1974). Small areas of Maungamaungaroa gley soils follow the main stream channels. Quickflow yield in 21 events during 1975 ranged from 0.3 mm to 65 mm, and total quickflow was 95 mm. A typical storm hydrograph is shown as Fig. 3. Total runoff digitised from the plotted hydrograph was 475 mm (TIDEDA total runoff was 481 mm). Quickflow was 20% of runoff and 7.3% of gross rainfall. Net rainfall for rye/clover pasture averages 80% of gross rainfall (Blake 1972); quickflow yield was 9.1% of net rainfall.

Infiltration data for the Manurewa soils have not been reported, but final infiltration rates for silt loam soils under permanent grazing are unlikely to exceed 20 mm/h (Musgrave and Holtan 1964). Rainfall intensity data for Mechanics Bay (15 km distant) from Robertson (1963) are shown in Table 2. It seems clear that rainfall rates frequently exceed the likely upper limit for infiltration rates in these soils, thus some storm runoff is probably produced by Horton overland flow, at least in events with return periods exceeding a few

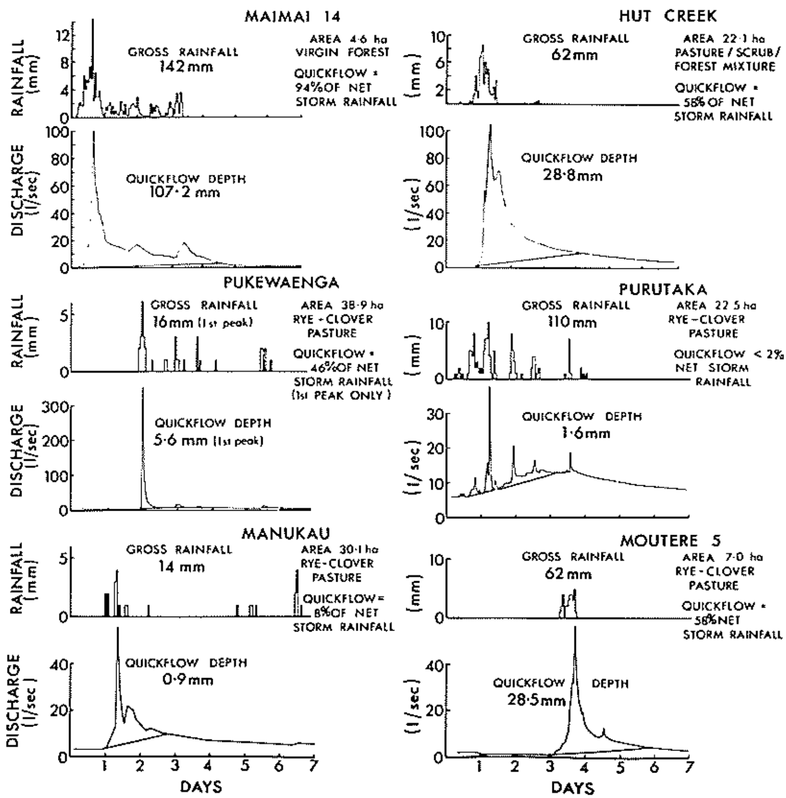

FIG.—3 Typical storm hydrographs from selected catchments. (Note the differences in scale between the various hydrographs.)

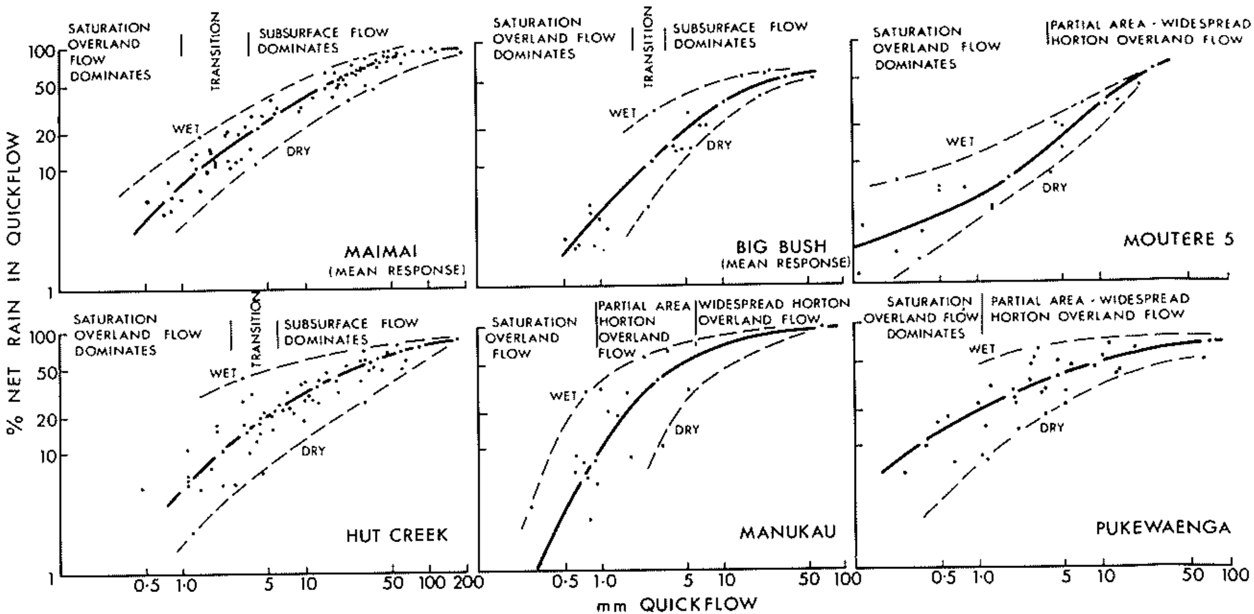

FIG.—4 Relationship between depth of quickflow and proportion of net rainfall yielded in quickflow. Heavy line shows average response to storm rainfall, envelopes for response in wet and dry antecedent conditions are dashed.

months. In more frequent events, much of the storm hydrograph is probably generated by saturation overland flow from the near–channel areas where poorly drained Maungamaungaroa gley soils occupy 6% of the catchment area (Orbell 1974).

Quickflow yields in the 21 events were calculated as a percentage of net storm rainfall for each event (Fig. 4) and averaged for three classes of event size. For 11 events yielding less than 1.0 mm quickflow, storm runoff averaged 9.8% of net rainfall. For nine events with yields between 1.0 mm and 5.2 mm, quickflow was 30% of net rainfall, and for the one large event, with 64.8 mm of quickflow, storm runoff was 98% of net storm rainfall. Thus it appears that in events with return periods of the order of weeks (< 1 mm of quickflow), the stormflow response could be produced from less than 10% of the catchment (in some instances from as little as 1% of the catchment area). Saturation overland flow on near–channel areas probably accounts for these events. For events producing 1–5 mm of quickflow, between 10% and 60% of the catchment area appears to contribute to storm flow, with an average of 30%. Horton overland flow on partial areas probably produces the stormflow in these events which have likely return periods of the order of a few months (nine such events in 12 months). Essentially all of the catchment appears to contribute to stormflow in larger events; this is consistent with generation of storm runoff by widespread Horton overland flow in view of the limited infiltration rates noted earlier. The pattern of stormflow response at Manukau is similar to that at Pukewaenga.

Pasture cover on pumice soils, Purutaka catchment

Purutaka catchment is one of three catchments in the Purukohukohu catchment study midway between Rotorua and Taupo (Fig. 2). The 22.5 ha catchment was originally covered in podocarp/mixed hardwood forest which was cleared for pasture development about 1920. The terrain is rolling to steep hill country covered by yellow–brown pumice soils which are derived from tephra deposits ranging in age from 1800 years to 13 000 years. The soils are from 160 cm to 220 cm thick, and are sandy loams and loamy sands, with lenses of pumice gravel and lapilli. The youngest Taupo Ash ranges from 10 cm to 70 cm in thickness over the older ash beds. The Recent tephra overlies a great thickness of variably silicified pumice, pumice breccia, and siltstone of Pleistocene age (M.W.D., 1971c, 1973b). Four events during 1975 yielded quickflow amounts ranging from 0.7 mm to 3.5 mm, and produced a quickflow total of 6.6 mm out of 393 mm of runoff (TIDEDA runoff total is 384 mm). A typical winter storm hydrograph is shown in Fig. 3. Only 1.7% of the total runoff (0.4% of the gross rainfall) was yielded in quickflow. Net rainfall for pasture cover here is estimated at 80% of gross rainfall; thus quickflow yield was 0.5% of net rainfall.

Infiltration data for the Purutaka soils are not available but Selby (1970, 1973) presents data for similar yellow–brown pumice soils at the Otutira catchments near Taupo. Final infiltration rates for sprinkler irrigation on grazed pasture were in the range 30–900 mm/h, and rates were up to 2400 mm/h on ungrazed pasture and scrub. Mean infiltration rates for pasture, scrub, and ungrazed grass were, respectively, 290 mm/h, 820 mm/h, and 1000 mm/h (Selby and Hosking 1971). On runoff plots of 4 m² area, Selby (1973) found that under pasture an average of 4.6% of gross rainfall was yielded in runoff, and an average of 0.8% of gross rainfall became runoff on ungrazed grass and scrub plots. Rainfall intensity–duration–frequency data are not readily available for Purukohukohu, but data from Robertson (1963) for Whakarewarewa (Rotorua) and Taupo, each 40 km distant from Purukohukohu are shown in Table 2. It seems clear that rainfall rates rarely exceed infiltration rates of well developed yellow–brown pumice soils that are moderately moist. As Selby (1973) and others have described, pumic soils display varying water repellent properties when dry; thus Horton overland flow may occur at times. When the soils are moist, however, infiltration rates probably exceed rainfall rates over most of Purutaka catchment, except for a small area (*c*.1 ha) where impermeable Waikokomuka soils have been formed by hydrothermal alteration of Taupo Pumice (Rijkse and Bell 1974).

Quickflow yields for three events yielding between 0.7 mm and 1.6 mm of runoff from between 29 mm and 68 mm of rainfall averaged 3.7% of net storm rainfall. Rainfall data for the larger (3.5 mm quickflow) event are not available. Storm runoff in the three smaller events could have been produced entirely by saturation overland flow from the small area (4.4% of the catchment) which has impermeable soils. The storm hydrograph shown in Fig. 3 reveals the response of the Purutaka catchment to storm rainfall when soils are moist. The initial short, sharp peak is almost certainly produced by saturation overland flow on the area of Waikokomuka soils, and the slow delayed rise to a broad second peak in the hydrograph must come from subsurface flow in the deep, permeable tephra. The small peaks superimposed on the delayed rise are the result of small rainfall events following the main rainfall event. The storm hydrographs from Purutaka show a strong resemblance to those of Freeze (1974, Fig. 4) and Hewlett and

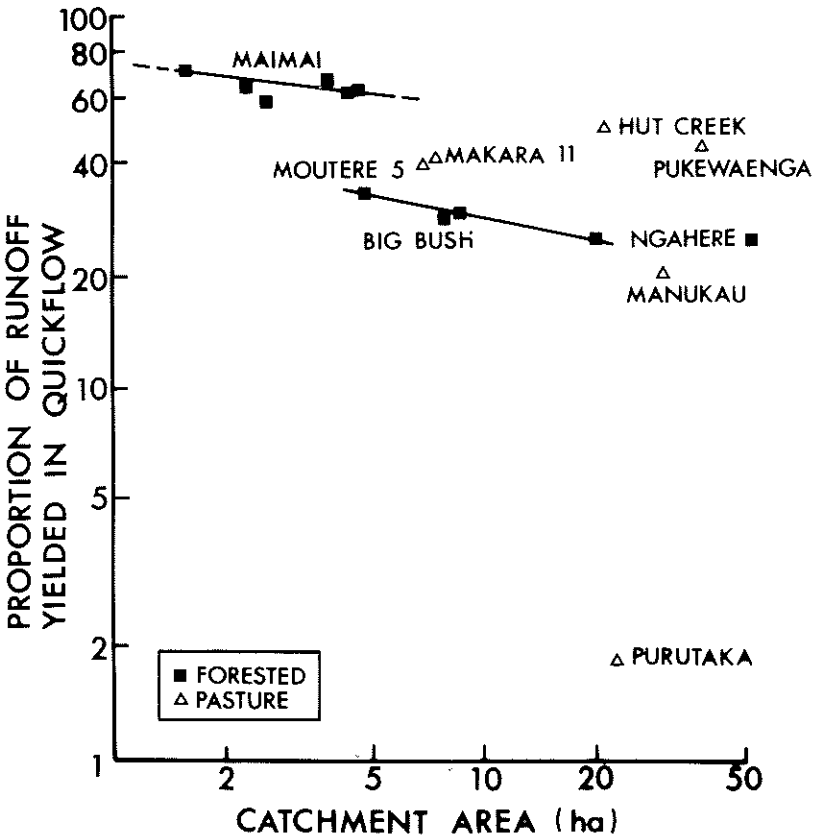

FIG.—5 Relationship between catchment area and proportion of runoff yielded in quickflow. No clear distinction on the basis of vegetation cover is apparent. The small quickflow yield of Purutaka is related to the hydrogeologic character of the underlying tephra deposits.

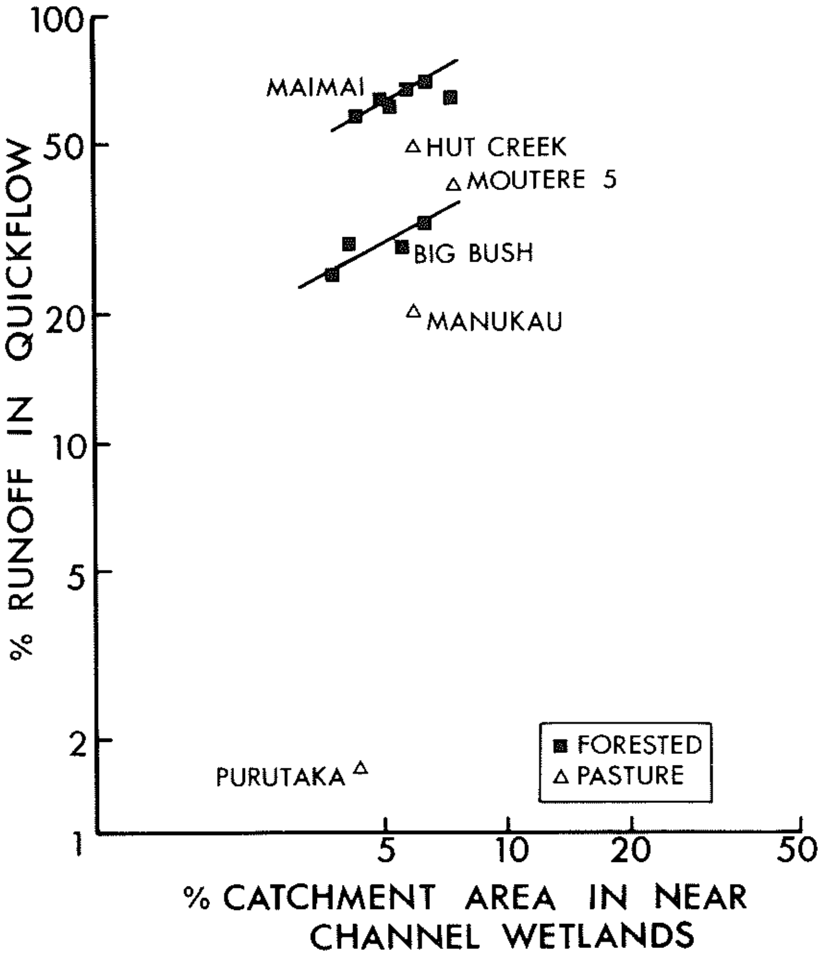

FIG.–6 Relation between proportion of catchment area in near-channel wetlands and proportion of runoff yielded in quickflow. Wetland areas for Makara 11, Ngahere and Pukewaenga are unknown; Moutere 5 estimated at 7.5%. Direct precipitation on the wetland area of Purutaka can account for all of the observed quickflow. For all other catchments, the size of the wetland area is insufficient to account for all quickflow.

Nutter (1970) from the Kimakia catchment in Kenya, and another Kimakia hydrograph shown by Dunne (1976, Figs. 7-38). The Kimakia catchment is also underlain by a great thickness of tephra, and it seems likely that the streamflow response exhibited by both Kimakia and Purutaka is typical of such regions. The bulk of the streamflow response to storm rainfall appears in the channel some days after the rainfall event; the slow rise and equally slow recession add support to the suggestion that this runoff is supplied via long subsurface pathways. The difference between the storm runoff yield at Purutaka and all other catchments studied is shown in Figs. 5 and 6. The much reduced response at Purutaka is mainly a result of the hydrogeology of the underlying tephra deposits.

Virgin red beech/mountain beech forest, Ngahere catchment

The 52 ha Ngahere catchment is a representative basin for the Kaweka hydrological region and consists of rolling–to–steep mountain slopes at a mean altitude of 1000 m; red beech forest on the lower slopes and mountain beech forest on the higher slopes cover 98% of the catchment, the remainder being in scrub. The soils have not been mapped in detail but are steepland yellow–brown earths derived from the underlying indurated sandstone and shale, with some yellow–brown pumice soils derived from the thin mantle of Recent tephra which is found in parts of the basin. Snow forms 5% of the mean annual precipitation (M.W.D. 1972a). Quickflow yield in 31 events during 1970 ranged from 0.1 mm to 141 mm, with a total quickflow yield of 416 mm. Total runoff digitised from the plotted hydrograph was 1686 mm (TIDEDA total runoff is 1661 mm). Quickflow formed 24.7% of total runoff and 15.8% of gross precipitation. Canopy interception at Ngahere is estimated as 23% for mountain beech and 16% for red beech (M.W.D. 1972a), but no measurement of understorey interception has been made. It seems likely that total interception would be in the order of 25% of gross precipitation, thus quickflow yield was approximately 21% of net precipitation.

Infiltration data for the steepland soils at Ngahere or for similar soils have not been reported. McDonald (1961) gives a figure for B–horizon permeability of Ruahine steepland soils of 157 mm/h (obtained using 10 cm diam. cores under 1 to 2.5 cm heads). Both the Ruahine steepland soils and the soils at Ngahere are steepland variants of central yellow–brown earths, and are derived from the same lithology; we assume that McDonald's figure is applicable to the Kaweka soils. Annual maximum rainfall data for various durations for Kaweka Forest and Makahu Saddle from 1969 to 1974 show that rainfall rates rarely exceed 30 mm/h for durations greater than 30 minutes and a 10–minute rainfall rate of 50 mm/h has a return period of about two years (N.Z. Met. Service 1969, 1970, 1971, 1972, 1973, 1974). Grant (1969) gives figures for the largest recorded event at Makahu Saddle; the maximum 20–minute rainfall rate was 53 mm/h. It seems from these figures that rainfall rates rarely exceed infiltration or hydraulic conductivity rates of the soils at Ngahere, thus storm runoff cannot be explained by the Horton mechanism.

Accurate storm rainfall amounts for Ngahere are unavailable as the only rainfall data are daily totals, thus the relationship between runoff event size and proportion of net rainfall yielded in quickflow cannot be investigated in detail. The average yield of quickflow is 21% of net rainfall (including all rainfall that did not produce storm runoff) thus it seems likely that, for individual events, quickflow yields will range from a few percent of net rainfall to several tens of percent of net rainfall. The steep slopes, thin permeable soils, and incised streams of the Ngahere catchment all fit Freeze's (1972, 1974) criteria for substantial contributions to storm runoff via subsurface flow at least in larger events. One of Freeze's simulations (1974, Fig. 18, Case B) with rainfall rate 12.5 mm/h for five hours, soil thickness 20 cm at the top of the slope and 100 cm at the slope base, slope length 33.5 m, slope 7.5% (4°), and saturated hydraulic conductivity 158 mm/h, produced a 12–hour yield of storm runoff via subsurface flow of 17% of precipitation. In all respects except that of slope, this model approximates the hillslope properties at Ngahere. Freeze (1972, Fig. 9), found that steeper slopes produced only a slight increase in stormflow yield via subsurface flow. It is clear that Freeze (1972, 1974) has provided theoretical support for the production of stormflow yields of the order of those at Ngahere by subsurface routes, provided soils are shallow and permeable, slopes are steep, and the channel network is incised. The storm runoff in smaller events is probably supplied mostly from saturation overland flow on near–channel areas; without more detailed rainfall and soil data, a more detailed apportioning of storm runoff to various mechanisms is not possible.

Pasture and scrub cover on steep hill-country, Makara catchment 11

The Makara catchments are on moderate and steep hill-country underlain by indurated sandstone and shale 8 km west of Wellington. Catchment 11 is a 7.4 ha steepland catchment which has been monitored since 1960 under pasture and scrub as a control for the Makara experimental programme (M.W.D., 1970b, 1972b). Twenty–five events during 1970 with quickflow yields ranging from 0.1 mm to 12.8 mm produced a total of 135 mm of quickflow, and a total runoff of 331 mm was calculated from the digitised plot (TIDEDA runoff total is 303 mm). Quickflow yield was 41% of runoff and 11% of gross rainfall. Net rainfall for pasture cover here is estimated as 80% of gross rainfall (Blake, 1972), thus quickflow yield is 13.9% of net rainfall.

Infiltration data for Makara steepland soils have not been reported but McDonald (1961) gives a figure for B–horizon permeability of Makara steepland soils of 148 mm/h. Rainfall intensity–duration–frequency data for Kelburn (about 10 km distant) from Robertson (1963) are given in Table 2. Even if McDonald's permeability data over-estimate infiltration rates by a factor of three, rainfall rates do not exceed infiltration rates except in very short duration events with return periods greater than 10 years. It seems unlikely that storm runoff from catchment 11 is generated by Horton overland flow except in events with return periods exceeding ten years.

Rainfall data other than daily amounts are not available on TIDEDA, thus the relationship between event size and proportion of net rainfall yielded in quickflow cannot be investigated in detail. Quickflow was 14% of net rainfall for the year, including rainfall events that did not produce storm runoff. In individual events, quickflow yields probably range up to a few tens of percent of net rainfall. Freeze's simulations (1974, Fig. 18, Case B.) support subsurface generation of stormflow yields of the magnitude found in catchment 11 at Makara on steep slopes with thin soils, and hydraulic conductivity of the order of 150 mm/h. Stormflow response at Makara 11 is probably similar to that at Ngahere.

Pasture cover on dissected hill–country, Moutere catchment 5

The Moutere group of experimental catchments is located about 30 km south–west of Nelson, on rolling–to–steep dissected hill–country underlain by weathered Moutere Gravels of early Pleistocene age. The catchments have been monitored since 1959 and various pasture improvement, forestry, and other treatments have been imposed (M.W.D., 1968b, 1970c, 1970d, 1970e, 1971d, 1972c, 1974). Catchment 5 is a 7 ha improved pasture catchment under grazing which is used as a control for the experimental programme. The soils are sandy loams and silt loams and range from 40 cm to 65 cm in thickness over weathered gravels in a silty–clay matrix (M.W.D., 1968b). Total quickflow in 1975 was 110 mm in 23 events with yields ranging from 0.1 mm to 28.5 mm. A typical storm hydrograph is shown in Fig. 3. Total runoff digitised from the plotted hydrograph was 277 mm (TIDEDA total runoff is 272 mm). Quickflow was 39.8% of total runoff and 9.9% of gross rainfall. Net rainfall is estimated as 80% of gross for rye/clover pasture (Blake, 1972), thus quickflow yield was 12.3% of net rainfall.

Published infiltration data for the soils at Moutere are not available but Duncan (1972) investigated the performance of a rainfall simulator at Moutere. Under a constant rainfall rate of 58 mm/h, mean steady-state infiltration (after about 1 hour) for six plots was 11.7 mm/h. On repeat runs after partial drainage of the soil, steady-state infiltration was reduced to 2–3 mm/h (Duncan 1972, pers. comm.). Zones of saturated or near-saturated soil exist in the valley bottom and also on the valley sides; these are related to the presence of less permeable clay loam subsoils shown on the soils map of catchment 5 (M.W.D., 1968b). These wet areas occupy about 5–10% of the catchment (Duncan pers. comm.). Annual maximum rainfall rate for ten minutes exceeded 30 mm/h in 13 out of 14 years, 30-minute rates exceeded 20 mm/h for all 14 years, 1-hour rates exceeded 15 mm/h in 12 out of 14 years, and 2-hour rates exceeded 12 mm/h in 12 out of 14 years. (N.Z. Met. Service, 1961–1974). It seems clear that rainfall rates for durations up to two hours frequently exceed infiltration rate, thus Horton overland flow is a feasible mechanism for producing storm runoff. Saturation overland flow can probably generate the storm runoff in events where 5% or less of net rainfall is yielded in quickflow, but Horton overland flow is the probable runoff mechanism in larger events.

Mean quickflow yield as a proportion of net rainfall for four size-classes of event is shown in Table 4, and for individual events in Fig. 4. The proportion of the catchment contributing to storm runoff varies from 3% to 43% depending on the depth of quickflow. Storm events yielding less than 1 mm of quickflow, and possibly many of those yielding up to 5 mm of quickflow, can be accounted for by saturation overland flow on less than 10% of the catchment area. In larger events, between 20% and 60% of the catchment must contribute to storm runoff; in view of the low infiltration rates, Horton overland flow appears the likely mechanism, provided that rainfall rates in these events exceed infiltration rates. Freeze's Case C simulation (1974, Fig. 18) approximates the conditions at Moutere 5 for rainfall rates less than infiltration rate; saturation overland flow dominates the storm hydrograph in this simulation (saturated hydraulic conductivity 15.8 mm/h, slope 7.5%, soil thickness 20 cm to 100 cm, rainfall rate 12.5 mm/h for five hours). The pattern of stormflow response at Moutere 5 is similar to that at Pukewaenga and Manukau.

TABLE—4 Quickflow yield as a proportion of net rainfall, Moutere 5.

Depth of quick-flow (mm)	*No. of events*	*Mean proportion of net rainfall yielded in quickflow (%)*	*Standard error of Mean*
0.05 – 0.99	12	3.3	±0.7
1.00 – 4.99	4	10.1	±4.2
5.00 – 14.99	4	22.6	±4.6
15.00 – 30.00	3	43.3	±9.4

Beech/podocarp/hardwood forest on dissected hill-country, Big Bush catchments

Four F.R.I. catchments ranging from 4.8 ha to 20.2 ha in size are located in Big Bush State Forest, 5 km north of the Hope Saddle (Fig. 2). The area is underlain by weathered early-Pleistocene Moutere Gravels. The terrain is moderately dissected, local relief is 100–250 m, slopes are up to 500 m long and range from 15° to 35°, averaging 28°. Soils are shallow, stony, podsolised, yellow-brown earths of the Hope Hill soils. The organic horizons of the soil average 8 cm in thickness. Mean annual rainfall at Kaka (4 km north-west of the catchments) is 1650 mm.

The hydrograph for each catchment from November 1 1975 to January 31 1977 was separated into quickflow and delayed flow, yielding the results shown in Table 5. The proportion of total runoff yielded in quickflow is related to catchment size (Fig. 5). smaller catchments yielding a larger proportion of total runoff in quickflow. The average proportion is 28.7%. Gross rainfall during this period was 2136 mm, thus 12.2% of gross rainfall was yielded in storm runoff. Interception studies indicate that net rainfall is 80% of gross rainfall, thus quickflow yield was 15.3% of net rainfall, including all rainfall which did not produce storm runoff. The average yield of quickflow from the four catchments as a percentage of net storm rainfall in 25 events ranging from 0.5 mm to 58 mm of quickflow is shown in Fig. 4. The proportion of net rainfall yielded in quickflow ranged from 2.2% to 58% and averaged 21%. Average proportions of net rainfall yielded in quickflow for five size–classes of event are shown in Table 6.

TABLE–5 Hydrograph separation 1.11.75-31.1.77, Big Bush catchments.

Catchment	*Area (ha)*	*No. of events producing > 0.025 mm quickflow*	*Quickflow depth (mm)*	*Total runoff*	*Proportion of Total runoff in quickflow (%)*
1	8.6	64	248	854	29.1
2	4.8	66	299	905	33.0
3	7.8	58	245	876	28.0
4	20.2	66	249	998	25.0

TABLE–6 Quickflow yield as a proportion of net rainfall, Big Bush catchments.

Depth of quick-flow (mm)	*No. of events*	*Mean proportion of net rainfall yielded in quickflow (%)*	*Standard error of Mean*
0.5 – 0.99	6	3.2	±0.5
1.0 – 4.99	8	11.3	±3.3
5.0 – 14.99	5	23.1	±3.5
15.0 – 29.99	4	50.5	±5.6
30.0 – 60.00	2	50.5	±0.6

Overland flow of the Horton type has not been observed on any of the catchments and is unlikely to occur because the infiltration rates of the litter and humus horizons probably exceed 1800 mm/h (0.05 cm/sec) and saturated hydraulic conductivity of the upper mineral horizons is probably of the order of 180 mm/h (0.005 cm/sec). Detailed rainfall intensity–duration–frequency data are not available for this locality but it is most unlikely that rainfall rates exceed infiltration rates; storm runoff is not produced by rainfall excess.

Direct rainfall on the channel and surrounding saturated areas adequately explains smaller events but stringent limitations on the proportion of the catchment contributing in this way are imposed by the terrain. The main stream channels are incised to depths ranging up to 3 m thus the maximum width of the saturated or near–saturated channel margin is severely constrained. The maximum area occupied by the channel and surrounding wetlands varies between 5% and 10% of catchment area (Fig. 6). Events yielding less than 5 mm of quickflow are probably produced mainly by saturation overland flow. An average of 50 events yielding less than 5 mm of quickflow have occurred during 15 months of record, so these events have return periods between 1 and 10 days.

Antecedent conditions affect the yield of stormflow from all storms, especially the smaller runoff events. One storm yielded 3 mm of quickflow from 10.8 mm of net rainfall when previous conditions were wet and, at the other extreme, a storm of 39.9 mm net rainfall yielded only 1.2 mm of quickflow. Because of these antecedent effects, an upper limit of quickflow yield (say 5 mm) cannot be a precise guide to the mechanism producing runoff in all storms. Nevertheless it is clear that for most storms yielding less than 3–5 mm of quickflow, the bulk of the storm runoff could be produced by saturation overland flow. Similarly, for most storms producing more than 5 mm of quickflow, at least half of the storm runoff must be generated by some mechanism other than saturation overland flow.

The steep straight slopes, shallow soils of high permeability, moderate slope length, and the incised channels which produce a convex break at the base of most slopes, present near ideal conditions for rapid subsurface flow. Freeze's simulations (1974, Fig. 18, Case A) provide theoretical support for the generation of stormflow yields of the magnitude found in the Big Bush catchments (Case A has the same parameters as Cases B and C referred to earlier, except that saturated hydraulic conductivity is 1584 mm/h; the 12–hour yield of runoff in Case A was 35% of rainfall). In larger storms, then, subsurface flow augments the runoff generated by saturation overland flow to produce the high proportion of net rainfall yielded in quickflow.

Beech/podocarp/hardwood forest on steep, dissected hill–country, Maimai catchments

Six catchments ranging from 1.6 ha to 4.6 ha in area are monitored by F.R.I. in Tawhai State Forest, 5 km north–west of Reefton (Fig. 2). The area is underlain by weathered early–Pleistocene gravels of the Old Man Group, covered by podsolised stony yellow–brown earths of the Blackball Hill soils. Soil profiles have organic horizons averaging 17 cm in thickness, and 60 cm of mineral horizon. Slopes are short ($<$ 300 m) and steep (average 34°), and local relief is in the range 100–150 m. Mean annual rainfall is 2610 mm; annual runoff is approximately 1500 mm (Pearce *et al*, 1976).

The hydrograph from December 1 1974 to February 28 1977 was separated into quickflow and delayed flow, yielding the results shown in Table 7. The proportion of total runoff yielded in quickflow is partly related to catchment area (Fig. 5) smaller catchments yielding a larger proportion of total runoff in quickflow. Catchment 15 departs substantially from the relationship between area and quickflow yield shown by the other five catchments. The proportion of runoff yielded in quickflow is also closely related to the proportion of the catchment in channel and surrounding wetlands, (Fig. 6); however, Catchment 5 departs from the relationship shown for the other five catchments. The near–saturated areas are easily recognised by extensive mottling and gleying of the upper soil horizons (McKie, 1977) which indicates that these areas are saturated or near–saturated for most of the year.

The average proportion of runoff yielded in quickflow is 64%. Total rainfall during this period was 5573 mm, thus the average proportion of gross rainfall yielded in quickflow is 39%. Interception studies indicate that net rainfall averages 73% of gross rainfall; quickflow was 53% of net rainfall, including all rainfall that did not produce storm runoff. These quickflow yields are very high

TABLE—7 Hydrograph separation 1.12.74-28.2.77, Maimai catchments

Catchment	*Area (ha)*	*No. of events producing > 0.025 mm quickflow*	*Quickflow depth (mm)*	*Proportion of total runoff in quickflow*	*Total runoff (mm)*
5	2.3	159	2210	64.6	3420
6	1.6	126	2224	70.8	3143
8	3.8	161	2201	67.2	3275
13	4.3	153	2508	62.1	4036
14	4.6	162	2286	62.9	3635
15	2.6	163	1590	58.7	2710

in comparison to figures for the eastern U.S. (Hewlett and Hibbert, 1967) and are considerably larger than yields reported from Oregon (Harr, 1977). The explanation lies in the combination of shallow highly permeable soils, on short steep slopes, and small catchment size. The storage available in the soils is small, water drains freely to the relatively impermeable substratum of weathered gravels, and near–saturation conditions in the soil profile are attained over large areas of each catchment during many storms.

The streamflow response to rainfall for 74 storms which produced an average of at least 0.5 mm of quickflow on the six catchments is shown in Fig. 4. Mean quickflow yields ranged from 0.5 mm to 187 mm from net rainfall depths between 9.5 mm and 209 mm. Quickflow as a proportion of net rainfall ranged from 4.3% to 99%, and averaged 44%. Thus the area of the catchment contributing to stormflow ranges from less than 5% to practically the whole catchment during some large storms with wet antecedent conditions.

Infiltration studies and laboratory tests show that the saturated hydraulic conductivity of the 17 cm thick organic horizon of the soil is 6100 mm/h (0.17 cm/sec), and that of the uppermost mineral horizon of the soil is 250 mm/h (0.007 cm/sec) (Webster, 1977). Robertson (1963) gives figures for the 2–year return period 24–hour rainfall and 20–year return period 24–hour rainfall at Reefton of 81 mm and 130 mm. Detailed rainfall intensity–duration–frequency data are available only for Greymouth, and these are given in Table 2. The saturated hydraulic conductivity of the upper mineral horizon exceeds the 50–year return period 10–minute rainfall rate, and the conductivity of the thick organic horizon exceeds it by more than an order of magnitude. Storm runoff is clearly not generated by the excess of rainfall rate over infiltration rate.

The proportion of the various catchments near saturation at all times ranges from 4.4% to 7.4% (Fig. 6) thus direct precipitation on these areas can account for runoff events which yield somewhat less than 10% of net rainfall in quickflow. The extent to which these areas expand during storms is not yet known but there seems little doubt that in the largest events (> 30 mm quickflow) storm runoff must be supplied mainly by subsurface flow.

The relationship between yield of quickflow and proportion of net rainfall in quickflow for six size–classes of event is shown in Table 8. Events producing less than 1 mm of quickflow can readily be explained by the saturation overland flow mechanism, since the average proportion of net rainfall yielded in quickflow (5.6%) closely coincides with the average proportion of catchments near–permanently saturated (5.7%). Events yielding 1 to 5 mm of quickflow (averaging 15% yield of net rainfall) are probably produced largely by saturation overland flow as it is not difficult to envisage a doubling of the wetland area in quite frequent events, and an upper limit of 15% for the saturated part of these catchments seems quite realistic. Runoff events with more than 5 mm of storm runoff have between 40% and 90% of the catchment contributing to stormflow. The steep slopes and high soil permeabilities combine to ensure that such proportions of the catchment could be saturated only during extreme events. During the 27 months of record, 45 events yielding more than 5 mm of quickflow have occurred, thus the smallest of these events have return periods of the order of two to three weeks. The bulk of the storm runoff in these events can be accounted for only by high rates of subsurface flow.

TABLE—8 Relationship between quickflow yield and proportion of net rainfall in quickflow, Maimai catchments

Depth of quick-flow (mm)	*No. of events*	*Mean proportion of net rainfall yielded in quickflow (%)*	*Standard error of Mean*
0.5 – 0.99	6	5.6	±0.6
1.0 – 4.99	23	15.2	±1.2
5.0 – 14.99	9	41.0	±4.7
15.0 – 29.99	16	56.5	±2.6
30.0 – 59.99	13	74.6	±4.4
60 +	7	87.1	±3.8

Freeze's (1972, 1974) simulation models again provide theoretical support for the magnitude of the subsurface stormflow yields found in the Maimai catchments. The hydraulic conductivity of the 17 cm thick organic horizons (6100 mm/h) lies between Freeze's 1974 Case A (1974, Fig. 18) of 1584 mm/h and his 1972 Case A (1972, Fig. 6) of 15 840 mm/h. The 1972 Case A figure was intentionally chosen as an unrealistically high figure well beyond the limits of reported data, and is more than 2.5 times greater than the conductivities found in the Maimai catchments. Freeze (1972) showed that soils of such conductivity (with other parameters the same as 1974 Cases A, B and C described earlier) could deliver almost 100% of storm rainfall via subsurface flow during a 12–hour period after the onset of rainfall. The 12–hour yield of rainfall for the soil with 1584 mm/h conductivity was 35%. The Maimai catchments have shallower soils with hydraulic conductivity four times that of Freeze's simulation which yielded 35% of rainfall, and slopes that are nine times as steep. All of these factors indicate that subsurface contributions to storm runoff should lie between 35% and 100% of storm rainfall. Freeze (1972, p 1280) concludes his treatment of slopes with convex profiles thus:

> "Clearly, maximum percentages of precipitation input will be delivered to the stream channels (by different mechanisms) for very high permeability soil veneers and very low permeability soil veneers. The minimum response will be recorded at the conductivity that just barely supports subsurface storm flow without allowing the formation of a near–channel saturated wetland."

The Maimai catchments are an outstanding example of near–maximum yield of storm runoff in larger events because of the presence of the highly permeable organic soil horizons.

Mixed pasture/scrub/forest vegetation, Hut Creek catchment

A single catchment of 22.1 ha underlain by high country yellow–brown earths of the Bealey Hill soils over indurated sandstone and younger coal measures has been monitored in Craigieburn Forest Park since 1969. The vegetation is a mixture of native and introduced grasses, manuka scrub, pockets of mountain beech, and trial plantings of various conifer species. The mean altitude of the catchment is about 850 m, thus both snow and partial ground frost are experienced during the winter months. The streamflow record from this catchment is of only moderate reliability and, although some periods of rainfall and runoff records are missing, it is nevertheless one of the longest and most complete hydrological records for the eastern Southern Alps.

Little is known of the hydrologic properties of the Bealey Hill soils, but Harvey (1974), working on similar soils of the Puketeraki set found saturated hydraulic conductivities in the range 360 mm/h to 0.36 mm/h with the upper parts of the profile normally having conductivities in excess of 100 mm/h. Gillingham (1964) reported final infiltration rates under sprinkler irrigation at Porters Pass of 85–105 mm/h. Hayward (1976) recorded infiltration rates in the Torlesse catchment ranging from 275 mm/h (on bare ground) to 2360 mm/h. Soons (1970) found that well–vegetated runoff plots underlain by high country yellow–brown earths of the Kaikoura set at Cass frequently yielded less than 1% of rainfall in runoff, and that the runoff yield was rarely more than 3% of rainfall.

In view of the above, it seems that infiltration rates exceed most rainfall rates and that storm runoff in this region is not produced by Horton overland flow. Rainfall records at Craigieburn Forest show that a 1–hour rainfall rate of 10 mm/h has a return period of about one year, and 15 mm/h has a return period of 3 to 5 years. Thus storm runoff from rain events with return periods measured in weeks or months must almost certainly be produced by some mechanism other than rainfall excess over infiltration.

Approximately seven years of hydrograph from Hut Creek (informal name) were separated into quickflow and delayed flow. Quickflow constituted 2570 mm or 49.5% of the total runoff of 5200 mm. The total rainfall for this period was approximately 10 500 mm, thus about 25% of gross rainfall was yielded as storm runoff over the whole period.

The relationship between depth of quickflow and the proportion of net rainfall yielded in quickflow is shown in Fig. 4 for 62 storms with reasonably reliable rainfall and runoff records, in which at least 1 mm of quickflow was produced. The 62 storms studied all occurred during a period of 54 months. Net rainfall was estimated as 80% of gross rainfall based on the following proportions of vegetation types and interception losses: 10% young pines, wide–spaced and pruned

TABLE–9 Relationship between depth of quickflow and proportion of net rainfall yielded in quickflow, Hut Creek catchment

Depth of quick-flow (mm)	*No. of events*	*Mean proportion of net rainfall yielded in quickflow (%)*	*Standard error of Mean*
1.0 – 4.99	23	14.1	±2.0
5.0 – 14.99	19	27.5	±1.9
15.0 – 29.99	8	46.3	±5.6
30.0 – 59.99	9	57.5	±3.8
60.0 +	3	70.3	±9.2

– 25% interception; 10% mountain beech – 35% interception (Rowe, 1975); 10% bare ground – no interception; 70% improved tussock and light manuka scrub patches – estimated 20% interception. The quickflow yield in these storms ranged from 1.0 mm to 167 mm, in storms ranging from 11.2 mm to 204 mm net rainfall; quickflow yield ranged from 5.4% to 81.7% of net rainfall and averaged 31.4%. Table 9 shows the relationship between depth of quickflow and proportion of net rainfall in quickflow.

The storm runoff generated in storms producing less than 5 mm of quickflow at Hut Creek can probably be adequately explained by the saturation overland flow mechanism. The channel area and gentler foot–slopes in this catchment are swampy at all times of the year and rarely occupy less than 6% of the total catchment area. Expansion of the saturated areas up to 15% of the catchment area during storms with return periods of several weeks to 2–3 months seems quite feasible, since the soils here are shallow (10 cm to 1 m thick) and overlie impermeable rocks. In larger storms, saturated areas could develop on gentler parts of the upper slopes and on some areas of bare ground. Non–saturated areas on better vegetated ground or steeper slopes would separate at least some of these saturated areas from the near–stream saturated zones; thus their contribution to storm runoff is not easy to assess. Even if all such areas could contribute to the generation of storm runoff, it is doubtful whether the total saturated area could approach 50% or more of the catchment area except in the very largest of storms.

It seems likely that in most storms producing more than 15 mm of quickflow, a substantial contribution from subsurface flow is necessary to provide the high proportion of net rainfall that is yielded in quickflow. The smallest of these events has a return period of about three months (20 such storms in 54 months). Some slopes in the catchment have convex profiles and others are concave, thus a complex pattern of both subsurface and saturation overland flow inputs is likely. The simulations of Freeze (1974, Fig. 18, Case B, and 1972, Fig. 11(b)) support the magnitude of the storm runoff yields, and the relative importance of different mechanisms.

Comparison of Hut Creek and Torlesse Catchments

The Torlesse catchment is situated on the eastern flank of the Torlesse range, 75 km north–west of Christchurch. Physical details of the catchment and the study programme are given by Hayward (1975). Although the catchment is in the same hydrological region as Hut Creek, it is larger, covers a wider range of altitudes, including a significant area in the alpine zone, and lacks the thin Tertiary cover that is present in parts of Hut Creek. Nevertheless, as one of the few catchments of the eastern Southern Alps where flow records are available, it forms an interesting comparison to the Hut Creek catchment.

Hayward (1976) grouped 29 storm events at Torlesse into four size–classes and calculated, as the ratio of storm flow to precipitation, the proportion of the 385 ha Torlesse catchment contributing to storm runoff. For storms less than 25 mm rainfall, the contributing area averaged 0.9%; for storms 25–50 mm, 50–100 mm, and greater than 100 mm, the contributing areas were, respectively, 1.8%, 9.7% and 21.2%. These estimated contributing areas are smaller than those outlined above for Hut Creek because:

(1) Hayward used gross rather than net precipitation as his denominator.
(2) Hayward modified both the Barnes and master recession curve techniques for hydrograph separation by extrapolating the recession limb back to beneath the hydrograph peak flow (Gray, 1970; Linsley *et al*, 1958), and thus underestimated the direct or storm runoff of each event.
(3) Both methods used by Hayward yield estimates of stormflow that are

between 50% and 70% of the stormflow calculated by the method we have used; on hydrographs with steep recessions, the estimates will be more similar than on hydrographs with flatter recessions such as those shown in Hayward's examples.

Despite the difficulties in comparison there is general similarity in the response of Hut Creek and the Torlesse catchment at least for the larger events. Taking events which Hayward assessed as producing at least 4 mm of stormflow, allowing net rainfall as 90% of gross (20% interception on the 50% of the catchment that is vegetated) and increasing Hayward's stormflow estimate by 50% to make the results more comparable to our own technique, we find the results shown in Table 10.

The three storms yielding less than 15 mm of quickflow had an average contributing area of 11.7% and the three storms yielding between 15 mm and 75 mm of quickflow had an average contributing area of 43.7%. These results are of the same order of magnitude as those found for Hut Creek. The three largest storms, which have return periods of at least one year, require a substantial portion of the hydrograph to be contributed by subsurface flow. The smaller storms, with return periods of the order of months rather than years, can probably be adequately explained by the saturation overland flow mechanism.

TABLE—10 Estimates of catchment area contributing to storm runoff, Torlesse catchment.

Date	*Net rainfall (mm)*	*Estimated Stormflow (mm)*	*Contributing area (%) (stormflow/net rain)*	*Hayward (1975) estimate of contributing area (%)*
29.4.75	65	24.6	37.9	22.0
6.6.75	82	6.9	8.4	5.0
18.8.75	50	7.2	14.4	8.6
15.4.74	111	54.0	48.7	29.0
20.1.75	95	11.7	12.3	7.5
11.3.75	165	73.5	44.6	27.0

DISCUSSION

It is apparent from the preceding data that Horton overland flow is not the principal mechanism generating storm runoff in many parts of New Zealand. For most of the catchments examined, storm runoff from frequent events (return period days or weeks) can be best explained by direct precipitation on saturated variable source areas. In larger events, with return periods of months or years, the mechanism producing storm runoff depends principally on the hydraulic conductivity of the soils, and on the form of the catchment hillslopes. Where hillslopes have concave lower slopes and especially where hydraulic conductivity of the soil is also low or moderate, partial area Horton overland flow, or widespread Horton overland flow dominates the storm hydrograph. The Pukewaenga, Manukau and Moutere 5 catchments are of this type. On steeper slopes, especially those where an incised channel produces a convex profile in the lower slope, and where soils are shallow and hydraulic conductivities are high, rapid subsurface flow dominates the storm hydrograph. The Maimai and Big Bush catchments are good examples, and the Ngahere and Makara 11 catchments probably behave in a similar fashion.

The behaviour of the Hut Creek catchment during larger events lies somewhere between these two extremes. The wide range of soil thickness–permeability–slope profile combinations in this catchment suggests that some parts of the channel are fed by saturation overland flow and others are fed by subsurface flow. In the largest events however, it seems likely that the contribution from subsurface flow is at least equal to that from saturation overland flow.

Purutaka catchment is a special case because of the unusual hydrogeologic properties of the underlying rocks. Small amounts of storm runoff are produced by saturation overland flow on near–channel wetlands, but the bulk of the response to storm rainfall occurs some days after the peak of rainfall. This delayed flow must be delivered to the stream via long subsurface pathways.

The very high yields of storm runoff from the forested catchments used in this study deserve comment. As noted earlier, the yields are explained by the combination of shallow, highly permeable soils and very steep slopes found in New Zealand hill and mountain forests. The hydrologic behaviour of these regions is dramatically different from that of the forested regions of the eastern United States where slopes are often longer and gentler, soils are normally much thicker with a smaller proportion of the profile in the very permeable organic horizons, and hydraulic conductivities are lower.

The low yields of rainfall in storm runoff which are typical of eastern United States forests (Dunne, 1976, Tables 7–5 and 7–13) are probably quite atypical of New Zealand hill and mountain forests. Therefore the maintenance or establishment of forest cover on New Zealand hill and mountainlands may not minimise storm runoff as is commonly believed. The extremely large stormflow response of the Maimai and Big Bush catchments is very strongly influenced by the presence of a thick, highly permeable forest floor or duff horizon. Without such a highly permeable soil veneer, storm runoff response would unquestionably deliver smaller proportions of storm rainfall less rapidly than at present. There are, of course, substantial hydrologic reasons for retaining or establishing forest cover in many hill and mountain areas, but extrapolation of the desirable low stormflow–response behaviour of eastern United States and European forests to many parts of New Zealand is probably an invalid justification for forest management and development in these regions.

This study has demonstrated the likely importance of the near–channel area in the production of storm runoff during small events in most parts of New Zealand. Important implications for the maintenance of the physical, chemical, and bacteriological quality of streamwater stem from this result. Land management operations such as cultivation, forest clearance, animal depasturing, and application of fertilisers, poisons, herbicides, and other chemicals in this region require strict control if the quality of storm runoff is not to be seriously degraded in small, frequent runoff events. Since the bulk of storm runoff in small events derives from near–channel areas, it is axiomatic that the principal sources of sediment transported in such events are located in the same region.

Lastly, the widespread importance of subsurface flow from large fractions of catchment area in generating storm runoff during large events indicated by our results has important implications for land and water resource management. Because runoff delivered via subsurface routes will normally be of higher physical and chemical quality than that delivered by surface routes, land management practices should be designed to maintain the large values of hydraulic conductivity that appear to be typical of many hill and mountainland soils in New Zealand.

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MOUNTAIN STREAM SEDIMENTS

J. A. Hayward

ABSTRACT

In the last few decades much information has been presented about sediment transport in flumes and lowland rivers but comparatively little is known about sediments in mountain streams. Because there are obvious differences in stream energy and sediment supply, mountain streams can be expected to behave differently from lowland streams.

Results and experience from mountain streams are reviewed and found to be sometimes confusing and conflicting. Results from five years study in the Torlesse stream catchment are presented as a contribution to a better understanding of mountain stream sediment behaviour.

Although the Torlesse stream catchment is commonly described as 'severely eroded', sediment yields have been found to be among the lowest reported values. Suspended sediments are estimated to contribute less than 10% of annual sediment yield. Bed load sediment yields are found to vary greatly between storms and are also found to be more dependent on sediment supply than on transport capacity of stream flow. A down-stream 'wave–like' movement of sediments is described.

Conventional bed load prediction equations over–estimate sediment yields by up to several orders of magnitude. Bagnold's (1966) concept of stream power and the proportion of stream power utilised in bed load transport is a more reliable approach to the determination of sediment yields.

INTRODUCTION

Each year our rivers deliver sediment to the oceans and in so doing play an important part in the evolution of our landscapes. But societies, like landscapes, are dynamic and in response to changing needs structures are built on or close to rivers for power generation, communications, flood control, water abstraction, etc. Sediment tends to be recognised only when it becomes a problem, for example in impairing the effectiveness of a structure. Alternatively, we may find that supplies of river sediments for use in construction and development are more limited than we assume. In both cases we need information about yield and supply rates of river sediments.

Prior to 1925 there were only a few isolated sediment measurements anywhere in the world. It is only within the last few decades that procedures for sediment measurement have been seriously considered. In that time much information has been presented about sediment transport in flumes and lowland rivers but comparatively little is known about sediments in mountain streams.

This paper reviews our knowledge of mountain stream sediments and presents some results from five years study in the Torlesse stream catchment, Canterbury, New Zealand.

Lowland Rivers

Total stream sediments are usually made up of three fractions: bed load, suspended load, and wash load. Although this separation is arbitrary there is general agreement that bed load moves by rolling, sliding or bouncing on or very near the bed. Suspended sediment is that fraction which is held in suspension by the upward momentum of turbulent flow. Wash load is the finest fraction of suspended load which settles only very slowly.

In lowland rivers, total sediment yields have been assessed either by direct measurement, by sampling, by estimating equations, or by correlation with suspended sediment measurements. Total sediment yield is usually regarded as bed load yield plus suspended load yield.

Direct Measurement

The simplest method of estimating sediment yields and mean transport rates is to trap material in a dam and periodically resurvey the change in volume of stored detritus, making allowances for variable trap efficiency. In this manner, Thompson (pers. comm.) has estimated the annual rate of sediment accumulation in Lake Roxbrough (Central Otago, New Zealand) at 330 tonnes per square kilometre of catchment. However, because surveys are limited to dam sites, and cannot provide information about relations between solid and fluid flows, a variety of alternative methods have been developed for the direct measurement of sediment load.

Sampling

In the United States, suspended sediments form the bulk of sediment yield. With sponsorship from several federal agencies a variety of suspended sediment samplers has been developed. The USDH48 depth integrating sampler is one such device (Federal Inter–agency Committee, 1948). In Europe, greater attention was paid to bed load measurement and a variety of pans, baskets, trays and boxes were developed. The Swiss Federal Authority Sampler (1939) is an example of one such device. The more recent Californian development of the Helley Smith (1971) pressure–difference sampler has provided a versatile instrument suited to a wide range of flow and sediment conditions.

In small catchments sediment yields have been monitored using multislot divisors, (Geib, 1933) and splitters (Brown *et al.* 1970; Parsons, 1955). For a larger catchment (470 km^2), Leopold and Emmett (1976) have described an ambitious programme to assess bed load sediments by using a 48 foot (14.6 m) gated slot across the river bed. This slot was fitted with an endless belt to remove sediment to a sump on the river bank.

Estimating equations

The general approach of estimating equations is to assume that the stream transports a capacity sediment load and that this is a function primarily of bed shear stress. If the size of the bed material, the bed shear stress, and the critical shear stress required to initiate movement are known, then a rate of bed load transport can be calculated. Unfortunately, results from various formulae are often drastically different. Consequently, formulae should be regarded as guides for planning and design. Those who need to use bed load formulae for estimating sediment yields in lowland rivers can refer to a number of critical reviews. (See for example American Society Civil Engineers, 1971; Graf, 1971; Herbertson, 1969; and Shepherd, 1963). A limitation on the use of these formulae for New Zealand lowland gravel rivers is that a majority have been developed for sand and silt bed systems. Bagnold (1966) is the only author who has attempted to describe a bed load transport equation without recourse to experimentally derived coefficients. His approach was to write bed load transport as a function of the rate of stream power expenditure.

Correlation with Suspended Sediments

Estimates of suspended sediments are in general more reliable than estimates of bed load, and easier to obtain. If a sediment rating curve can be established (flow rate v sediment concentration) information about total suspended sediment yields can be derived from information about stream flows. To estimate total sediments it is often assumed that bed load is a fixed percentage of suspended load. Although this may vary from 2% to 20% (Gregory and Walling, 1973) a figure of 10% is frequently adopted, (Maddock and Borland 1951).

These methods and procedures have been used over the last few decades to provide much information about stream sediments from lowland rivers. In the same period little attention has been paid to mountain streams, where, because of obvious differences in stream energy and sediment supply the pattern of behaviour should be different.

Mountain Rivers

Results from the comparatively few studies of sediment in mountain streams are sometimes confusing and in conflict.

Table 1 presents mean annual sediment rates as estimated from reservoir surveys in a range of New Zealand catchments. These results suggest that the models of Langbein and Schumm (1958) and Fournier (1960) in which sediment yield is regarded as a function of climate are inadequate for New Zealand's steep and mountain lands. Likewise the models of Strakov (1967) (sediment as a function of relief) and Corbell (1964) (sediment as a function of climate and relief), are also inadequate. Although much information is concealed in the results presented in Table 1, they do provide a valuable first step toward the regionalisation of sediment yields.

TABLE—1 Mean annual sediment rates as estimated from dam and reservoir surveys for some New Zealand mountain catchments.

Location	*Yield* $m^3\ Km^{-2}\ yr^{-1}$	*Source*
Frazer Central Otago	30	1
Opuho	70	1
Opihi	80	1
Tangawai	220	1
Otaki	1000	1
Mangahao	1600	1
Tuku Tuki River, Folgers Lake	1500	1
Lake Tutira	1700	1
Waipoa	6500	1
South Eastern Ruahines	1000-5000	2
North Westland	55	3
Nelson	5	3

1. Presented by Mosley (Table 8)
2. Mosley 1977
3. O'Loughlin *et. al.*, 1978

From the studies reported by Leaf (1966), it can be assumed that annual sediment yields will vary by an order of magnitude or more about these mean values. Variation in sediment production within any one season can also be expected in response to variations in stream flow and sediment supply. For example, Nanson (1974) and Sunborg (1956) working in Canada and Sweden both found that sediment concentrations increased dramatically with seasonal peak discharge but that concentrations per unit discharge declined after the peak of seasonal snow melt. This decline was attributed to a reduction in sediment

TABLE–2 Suspended sediment yields for seven storms Torlesse Stream Catchment.

Date	Time	Concen-tration ppm	Flow $m^3 sec^{-1}$	Yield kg	Storm yield kg	Bed load yield for same period	Bed load yield for storm	Total sediment yield	s.s as % of total sediments	s.s as % of total storm sediment	Stage
12.8.73	0700 1000	225	0.270	655		84		739	89		Rising
	1800 2000	114	0.350	287		67		354	81		Rising
	2330 0130	276	0.370	735		107		842	87		Peak
					1677		15000	16700		10	
20.8.73	1500 1600	61	0.240	53	53	10	75	128	84	41	Rising
29.8.73	0430 0530	92	0.400	132		74		206	64		Rising
	0515 0715	3542	0.450	1147		478		1625	70		Rising
	1400 1500	365	0.700	919		3800		4719	20		Peak

	1800 1900	1688	0.540	1337		5715		7052	19		Falling
30.8.73	1400 1500	131	0.570	268		225		493	54		Falling
					2770		74000	76770		4	
7.10.74	1800 1900	40	0.500	72		50		122	59		Peak
	1900 2100	24	0.470	41		20		61	67		Falling
					123		150	273		45	
29.1.75	0900 1200	620	0.600	4017		936		4953	81		Peak
					4017		23000	27000		15	
Totals for 5 storms					8460		112225	120865		7	

supply. Studies by Takei *et. al.*, (1975) and Hayward and Sutherland (1974) have also shown variations in bed load sediment yield and suggest that these variations may be more dependent on sediment supply than on stream flow. Milhous and Klingeman (1971) found that bed load transport per unit water discharge increased after peak flow. They attributed this increase to a breakdown in bed armour which allowed sediments to be supplied from the stream bed.

Many studies of suspended sediments on lowland rivers have shown a hysteresis loop concentration associated with rising and falling stage. However, Nanson's (1974) results suggest that a series of sub–parallel rating curves may more accurately describe this relationship for mountain streams. He concluded that variations in this relationship were due to variations in sediment supply.

Although sediment estimating equations have been used to assess annual yields and transport rates, a majority of these have been developed from laboratory flume data. Field verification has generally been limited to sand and silt bed rivers. Kellerhalls (1972) reported that in steep gravel–bed rivers in Canada, published formulae give inconsistent and widely divergent results. Hayward and Sutherland (1974) showed that even the 'most appropriate' formulae overestimated sediment yield by up to two orders of magnitude.

The relative proportions of bed load and suspended load in total stream sediments are much more variable than for lowland rivers. For example, Jarocki (1957 quoted by Gregory and Walling, 1973) estimates that bed load accounts for 70% of the total sediment from alpine streams. Kellerhalls (1972) and Leaf (1966) both confirm that bed load makes a significant contribution to total sediments in the mountain rivers of Alberta and Colorado. On the other hand the studies of MacPherson (1971) and Nanson (1974) show contrary results. At Bragg Creek, Alberta, MacPherson found that bed load accounted for less than 1% of total annual sediment yield. An analysis of Nanson's data for Bridge Creek, Alberta, confirms this finding. Klingeman and Milhous (1970) showed that for Oak Creek, Oregon, the proportion of bed load in total stream sediments varied with flood magnitude. At flood discharges corresponding to the mean annual flood, about 25% of sediment yield was transported as bed load. For discharges with a return period of about once in 20 years 40% or more of sediment was transported as bed load. While it may be possible to explain such widely divergent results in terms of catchment size, glacial history, lithology, flood magnitudes and the like, they serve for the moment to illustrate the variability which is associated with mountain stream sediments.

THE TORLESSE STREAM CATCHMENT STUDY 1972-1977

The Torlesse stream drains a 385 ha mountain basin in the Torlesse Range. Altitude rises from 760 m to 2000 m within about 6 km. At the stream's outlet, a controlled section incorporating a vortex tube sediment trap was built to monitor stream flows and sediment movement. Details of the Torlesse programme and vortex tube sediment trap have been published by Hayward (1975), and Hayward and Sutherland (1974).

Methods of Measurement

Suspended sediments were monitored with a DH48 suspended sediment sampler. Bed load sediments were trapped and weighed after delivery through the vortex tubes (Fig. 1). The vortex tube trap has proved to be a most satisfactory if physically demanding method of estimating bed load movement from this catchment. However, two observations are relevant to others contemplating using this method.

The first concerns the operating range of the trap. From his investigations of the performance of vortex tubes, Robinson (1962) suggested that the Froude number of flow over the tubes should be in the order of 0.8. However, he also noted that Froude number had little effect on trapping efficiency at stage heights of less than 1.5 times the width of tube opening. This suggested that the Torlesse trap might be effective up to stage heights of 0.4 m. During floods of April 1974 and April 1978 (recurrence interval between 1:5 – 1:20 years) trap efficiencies were observed to be impaired at stage heights greater than 0.30 m. It is evident that the vortex tube trap can provide valuable information about sediment movement in more frequent events but is less useful in large low-frequency floods.

The second observation concerns the sediment discharge rate through the vortex tubes. From experience during the study period sediment rates in excess of about 200 kg min^{-1} either cannot be handled in the work pit or cannot be discharged through the vortex tubes. Within the trap's operating range such rates were found rarely and for only very brief periods. However, it will be shown later that the sediment yield from the Torlesse stream is controlled by its supply and that 'capacity' loads have not been recorded. From Fig. 11 it can be inferred that with 'capacity' load the trap would be unable to function at stage heights in excess of about 0.12 – 0.16 m. This would suggest that if sediment were freely available to the Torlesse stream, this trap would have been able to monitor sediments from only 10% – 20% of all flood events.

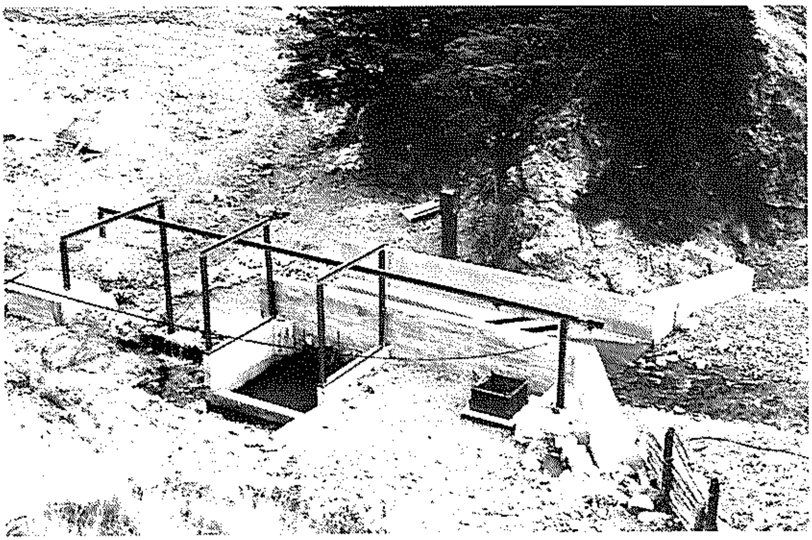

FIG.—1 The Torlesse stream vortex tube sediment trap. Two open tubes are embedded in the floor of the control section and lead through the true-left side wall into the work pit. Bed loads are retained in the steel basket, periodically weighed and returned to the stream via the monorail.

Suspended Sediments

Suspended sediments were found to be transported for brief periods during some storm events. This response in the Torlesse stream is in marked contrast to the adjacent Kowai River which drains a larger catchment and in which suspended sediments are transported throughout each storm. Table 2 summarises suspended sediment yields from five storm events for which reliable information is available. Suspended sediment sampling was discontinued in 1975 when it was found that suspended sediments were only a minor fraction of total annual sediment yield but their determination required a disproportionate amount of time and effort.

Suspended sediments were observed to move through the sediment trap in 'clouds' or 'waves'. These 'clouds' were characteristically of one to three hour durations. In most storm events there were prolonged periods during which suspended sediment was not transported and the water was clear enough to observe bed load movement. Field observations showed that the passage of each cloud of suspended sediment was related to a sudden influx of sediment into the stream channel; for example in the case of a bank collapse.

Table 2 shows that when suspended sediments are being transported they may account for up to 90% of sediment yield. However, because suspended sediments are transported for only a small proportion of total storm time they account for much less of total storm sediment yield. Table 2 also shows that the percentage of total sediments, that is suspended sediment, is greatest in small events. It will be shown later that these small events contribute little to total annual sediment yield. These results support the findings of Leaf (1966) and Kellerhalls (1972) that suspended sediment can be a relatively minor component of total sediments in some mountain streams.

The time distribution of suspended sediment was similar to that reported by O'Loughlin *et. al.* (in press) in that the highest concentrations occurred on the rising limb. Field observations suggest that fine material perched on stream banks was entrained on the rising stage. On a falling stage such sediments were generally unavailable to the stream.

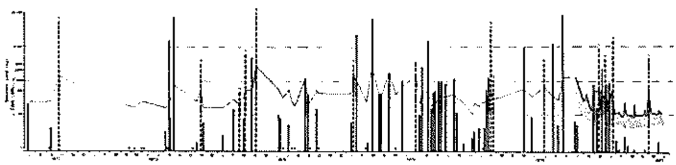

FIG.—2 A chronological sequence of bed load sediment yields (an associated peak flow rate) Torlesse stream catchment 1972-1977. (Sediment yields shown as bars, peak discharge shown as hatched line.)

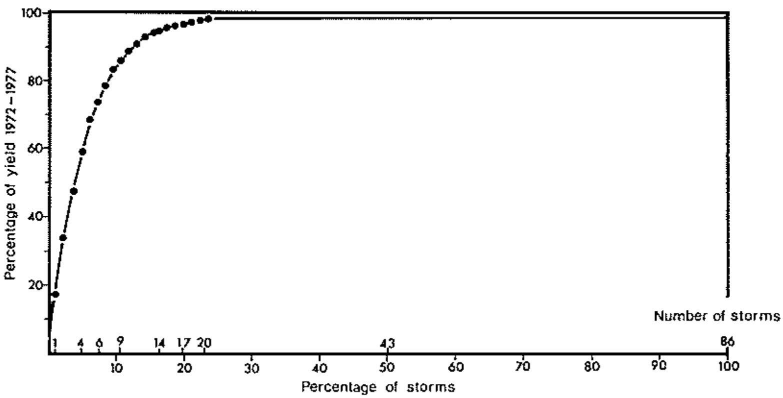

FIG.—3 Accumulated bed load yields Torlesse stream catchment 1972-1977.

TABLE—3 Accumulated bed load yields, Torlesse Stream catchment 1972-1977.

Peak flow class m^3/sec (depth m)	*Bedload sediment yield (tonnes)*				*No. of storms*	*Min*	*Max tonnes*	*Mean*
.150 – .200 (0.10m)	0 0.058 0.155 0.005	0 0.025 0.02 0.03	0 0 0	0 0.107 0.04	14	0	.155	0.03
.201 – .250 (0.115m)	0 0.015 0.014	0 0 0.032	0 0 0.002	0.075 0.44 0.034	12	0	.44	0.05
.251 – .300 (0.13m)	0.042	0.136	0	0.050	4	0	.136	0.06
.301 – .400 (0.15m)	0.01 0.84 0 0.015 1.20	0.078 0.20 2.15 0.40 1.00	0.002 0.54 0.58 0 17.6	1.0 0.085 0.405 1.0	19	0	17.6	1.43
0.401 – 0.500 (0.165m)	29.6 1.26 0 4.0	0.080 0.20 0.40	0.06 2.5 3.0	2.5 0.15 15.0	13	–	29.6	4.52
0.501 – 0.600 (0.18m)	6.0 10.6	4.7 0.50	0.095	27.8	6	0.095	27.8	8.28
0.601 – 0.800	0.05	1.0	0.05	73.6	5	0.05 73.6 15.64 (0.20m)	3.5	
1.00 – 1.10 (0.23m)	27.0				1			
1.21 – 1.30 (0.245m)	50.0	0.9	91.2	.065	4	.065	91.2	35.5
1.41 – 1.50 (0.26m)	5.0				1			
1.75 (0.28m)	1.57				1			
2.8 (0.35m)	72.0				1			

Bed Load Sediments

Bed load sediment yields from 81 storm events between 1972 and 1977 are shown in Fig. 2 and Table 3. The total yield of sediment amounts to 564 tonnes. The average annual sediment yield was 115 tonnes or 30 tonnes per square kilometre of catchment. Fig. 3 shows that a majority of this sediment was delivered in only a few events; 10 storms produced 90% of the 5–year total. As these storms took place over a total of 17 days, 90% of the sediments were delivered in 1% of total time. Table 3 shows sediment yields from the 81 recorded storms grouped into 12 storm classes, based on peak flow.

Yields

Reliable bed load data are sparse and it is therefore possible that the results presented above, like those presented by Leopold and Emmett (1976), have greater value than indicated by analyses or interpretations which are possible at the time of writing. The estimated specific annual sediment yield of 30 tonne km^{-2} is approximately 15 m^3 $km^{-2}yr^{-1}$ (density 2.0 g cm^{-3}). This indicates that erosion rates in the Torlesse catchment are among the lowest reported values (see Table 1). If this result is taken at its face value it is indeed surprising. Fig. 4 and detailed surveys by Saunders *et. al.* (pers. comm.) suggest that over 50% of this catchment is in a 'severely eroded' condition. However, a re-examination of the eroded areas shows that the coarse scree and block fields are much more stable than the depleted condition of their vegetation might at first suggest.

FIG.–4 The Torlesse stream catchment.

This contemporary rate of erosion can be compared with a long–term or natural erosion rate. A paleolandscape was reconstructed by the simple but subjective method of extrapolating the present day contours. The volume of material removed from the paleocatchment to create the present catchment was distributed over an assumed 2.5 million year period of landscape evolution. Although very crude, this method indicated a long–term erosion rate in the order of 400 m^3 km^{-2} of catchment per year. This suggests that erosion rates measured between 1972 and 1977 might be at least an order of magnitude less than long–term or natural rates of erosion.

Although the study period was reasonably representative of both weather and climate it did not include an 'extreme' event. In April 1978, a storm of return period thought to be about 20 years, delivered about 800 tonnes of sediment in 70 hours. Although this event is not included in the data presented here, it is mentioned to indicate the significance of low frequency events. Wolman and Miller (1960) have suggested that the greatest quantity of material is moved by small, frequent events. Our experience suggests that the frequency–magnitude concept is inappropriate to coarse gravel mountain systems. Similar views have been expressed by O'Loughlin (1978) and others.

If the study period 1972–1977 is extended to include the event of April 1978, specific sediment yields increase from 30 tonnes km^{-2} yr^{-1} to about 60 tonnes km^{-2} yr^{-1}. Although it is a matter of some speculation, it is possible that if the study could be extended for a much longer period to include several low frequency events, contemporary erosion rates might be found to be of the same order as the crudely estimated long–term or natural rates of erosion. Be that as it may, these results question the validity of the generally accepted view that contemporary erosion has increased stream sediment yields in this mountain catchment.

Variability

While Fig. 2 shows that the largest sediment yields are generally associated with highest peak flows, there are exceptions. For example, storms on 8 September 1976 and 11 October 1976 both produced peak flows of 1.3 m^3 s^{-1}. Sediment yields were 91 000 kg for the first storm and 65 kg for the second. Results such as these give a new perspective to the movement of sediment from this catchment.

Fig. 2 gives the impression of a periodicity of sediment movement which is only partially dependent on flood magnitude. Table 3 shows that there is great variability in sediment yield from floods of similar peak discharge. While part of this variation can be explained in terms of hydrograph shape and duration of flood flows, a more important explanation involves the availability of sediment.

During the study it became clear that sediment yields were strongly influenced by the amount of sediment held in storage in the stream channel. In turn, this was influenced by the stability of a restricted area of upstream riparian land. In 1974 the programme was extended to monitor changes in the storage of channel sediments.

Although streams such as the Torlesse have been generally described as mountain torrents, this description is inappropriate. The Torlesse stream channel is a sequence of pools and riffles. Sediment is stored within the pools and released by floods. A series of cross sections was established throughout the channel to monitor changes in the volume of detritus held in storage. Fig. 5 indicates the changes which took place in a pool 75 m upstream from the sediment trap. Three features should be noted about Fig. 5. Changes in storage are expressed as changes in the cross sectional area with respect to a mean bed level. Values below the mean bed line represent scouring and levels above represent aggradation. An alternative view would be to consider everything above the lowest recorded levels as sediment held in storage.

FIG.–5 Changes in cross-section area, pool 2A, Torlesse stream catchment.

Changes in cross sectional area can be rapid. The general behaviour was that pools scour on a rising stage and refill on a falling stage. However, many changes went unrecorded during a storm because of the difficulties of monitoring sediment yields and channel conditions at the same time. While Fig. 5 underestimates the frequency of channel change it does indicate periodic increases in the volumes of stored sediments. These were found to be coincident with variability in recorded yields at the trap.

A clearer pattern emerges when changes at all cross sections are plotted on the long profile of the Torlesse stream. Fig. 6 (a, b, c) shows the measured and inferred changes which took place over a 30 hour period in April 1975. This figure gives support to field observations of sediment moving wave–like down the stream channel. In this storm the bulk of the 'wave' did not reach the sediment trap and the recorded yield of 1500 kg was low for a peak discharge of 1.90 m^3 s^{-1}. Another gravel wave was observed to pass through the Torlesse stream channel in two events between December 1976 and March 1977. This wave travelled 3.5 km in 20 – 24 hours, say 0.15 km hr^{-1}.

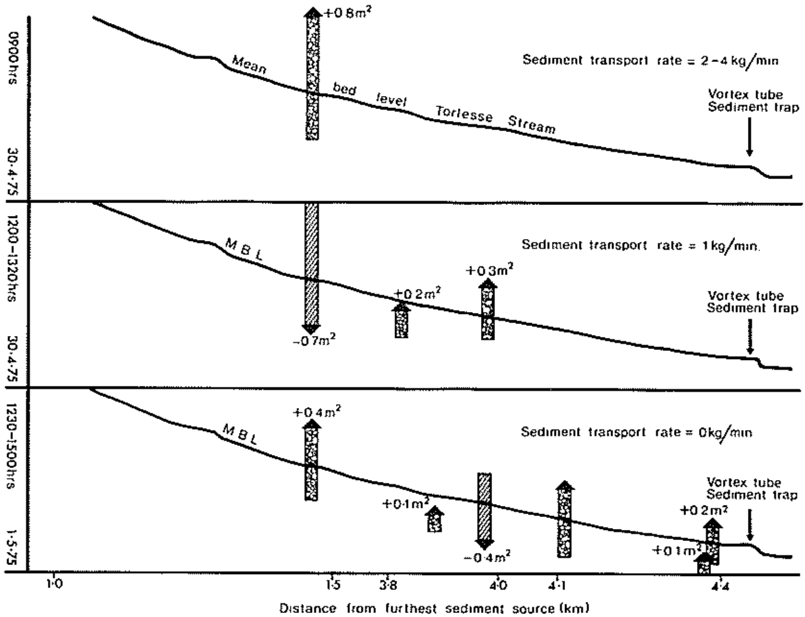

FIG.—6 Changes in cross-sectional areas, Torlesse stream catchment, 30 April and 1 May 1975.

In the five year study period there was a total of about 650 hours of sediment transport time. This is an average of about 10 hours per month. Based on a velocity of 0.15 km hr^{-1} the average residence time for gravel once deposited in the Torlesse stream channel would appear to be 1 – 3 months.

Experience during the study period suggests that it is both the presence (or absence) and location of these sediment 'waves' which is the main determinant of storm sediment yield.

In presenting results from the Hirudani experimental catchment, Japan, Ashida *et al.*, (1975) show variations in stream bed levels similar to those described here. Although those authors do not present results to show variability in sediment yields, it is reasonable to conclude that such variability does exist.

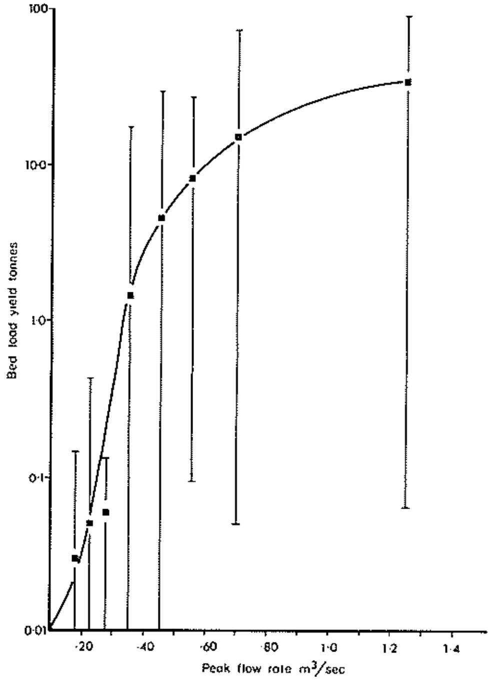

FIG.—7 Torlesse stream catchment range of sediment yields (minimum, mean and maximum) and peak flood flows, 1972-1976 (logarithmic scales).

While Fig. 2 and Table 1 show a wide range in sediment yield from floods of comparable peak discharge the mean values show an orderly progression with peak flow rate. The line which can be fitted by eye through these mean values represents the average behaviour of the Torlesse stream channel (Fig. 7). Using the terminology of Blench (1957) this relationship can be considered as a first approximation of regime for the Torlesse stream catchment. If further study can confirm this relationship and extend it to higher flow rates and sediment yields then future long–term sediment yields could be predicted from information about flood frequency.

For a particular storm, sediment yield will be determined by the amount of sediment held in storage. When stream bed levels exceed mean bed level, yields will be high and the stream will be in a condition of oversupply. Conversely, the stream will be undersupplied when bed levels are below mean bed level (Fig. 8).

Size

The smallest bed load particles retained in the trap were coarse sands. These were observed in all flows capable of transporting bed load. The largest particles to be ejected by the vortex tubes were in the order of 0.4 m (long axis) and 20 kg mass. At flow depths in excess of 0.30 m (2.0 m^3 s^{-1}) boulders in excess of 0.5 m and 40 kg mass were observed to roll or slide over the vortex tubes.

Fig. 9 illustrates the variability in particle size distribution throughout a storm. Because only a limited number of samples were retained for size analysis, Fig. 9 probably understates the variation in the proportions of bed load sediments during storm events. Observations of bed load movement through the sediment trap confirm these significant changes in the size distribution of sediments. While

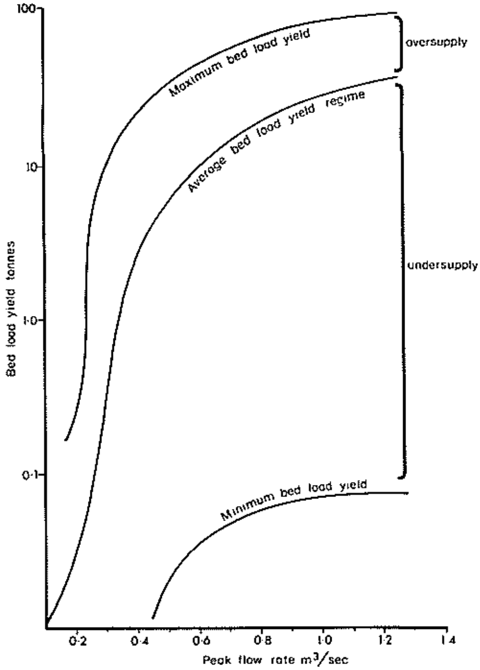

FIG.—8 Torlesse stream catchment, sediment yields 1972-1976 (logarithmic scales).

the reasons for these changes are the subject of further study it is probable that they reflect changes in the upper catchment and channel. For example, the collapse of a stream bank may cause an influx of sediment to the stream channel. The preferential movement of fines results in a shift in size distributions of sediment retained at the trap. Coarse sediments may be stored or detained in pools and subjected to a degree of sorting. The subsequent release of these coarse gravels from a pool (or pools) may result in substantially coarser material being discharged through the trap.

Comparison with Estimating Equations

Fig. 10 shows the distribution of bed load transport rates (kg min^{-1}) through the 3 m wide sediment trap for a range of flow depths. Upper and lower envelope curves and a line of best fit are also shown. The upper and lower curves are fitted by eye.

Fig. 11 compares bed load transport rates estimated by five prediction equations with measured rates. Fig. 12 shows the range of possible transport rates depending on the value used for slope in the estimating equation. In each case the lower estimates are based on the slope of the water profile through the pool (0.025). The upper estimates represent transport rate when the pool–riffle features are 'drowned' by water and sediment and the water surface slope approximates valley slope (0.067). For both measured the computed transport rates particles sizes were, $d_{50} = 0.025$ m, $d_{90} = 0.050$ m. Many authors have urged caution in the use of bed load prediction equations. These results confirm Kellerhall's (1972) view that conventional estimating equations are inappropriate for estimating sediment yields from mountain catchments.

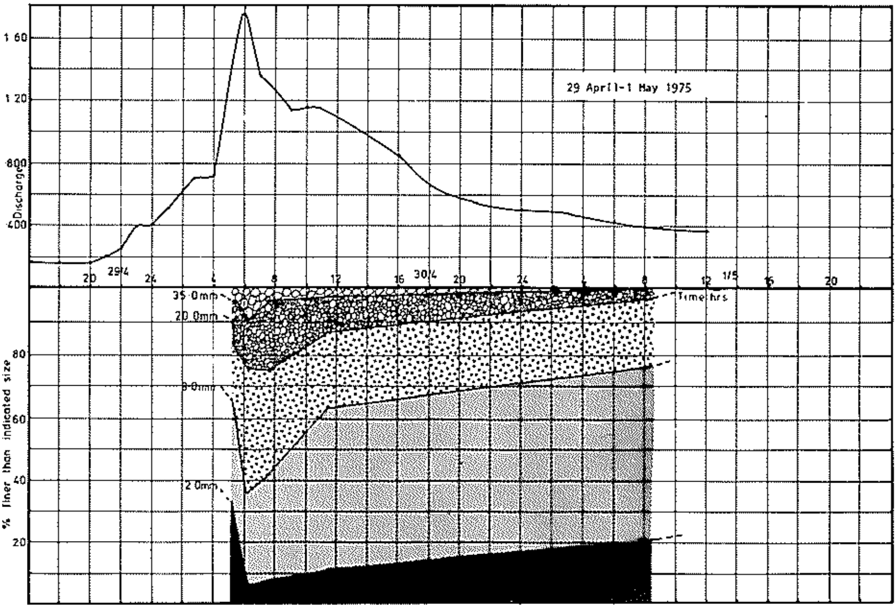

FIG.—9 Variability in distribution of bed load particle sizes throughout one storm Torlesse stream catchment.

Shield's formula has been found to over–estimate yields in a wide range of flume and field conditions (White, 1973). In this study the estimates from Shield's formula were consistently three orders of magnitude higher than the measured values. The best agreement came from Engelund and Hansen's formula (quoted by White *et al.* 1973) which gave estimates in the same order of magnitude as the measured values. However, as this formula was developed in a laboratory flume using sand sized particles, this agreement should be regarded as fortuitous.

It was noted earlier that most estimating equations assume that a capacity load

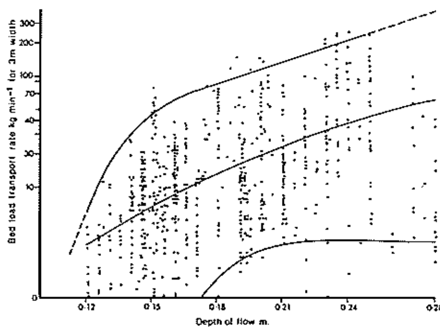

FIG.—10 Sediment transport rates at the Vortex tube sediment trap. Upper and lower envelope curves fitted by eye. Line of best fit for depths between 0.12 m and 0.28 m $y = 4819\ D^{3.5}$ where y = sediment transport rate (kg min^{-1}) D = depth of flow (m).

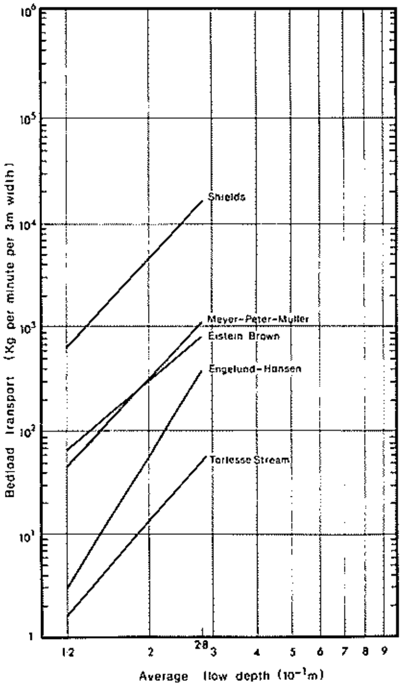

FIG.—11 Bed load sediments recorded at the Torlesse stream sediment trap *v* those estimated by some conventional estimating equations.

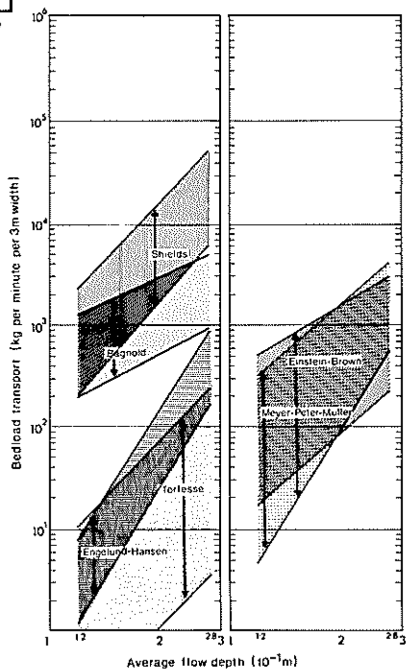

FIG.—12 Bed load transport rates estimated by some conventional equations. Estimates for each equation are within the slope ranges 0.025 and 0.067.

is transported and that this is related to bed shear stress. These assumptions are not valid for the Torlesse stream and therefore the use of such estimating equations is inappropriate for this or similar mountain catchments.

Stream Power and Sediment Yield

Bagnold (1966) is the only author to describe theoretically bed load transport without recourse to experimentally derived coefficients. In Bagnold's approach the rate of bed load transport should be a function of stream power. Stream power is the product of mean flow velocity and bed shear stress. That is, stream power is the product of water density, velocity, depth and slope. When sediment is freely available, Bagnold has suggested that at low values of stream power transport rate increases rapidly with increases in stream power. However, at higher values of stream power further increases in bed load transport are a direct and linear function of stream power. This hypothesis was tentatively confirmed by Leopold and Emmett (1976) using measured bed load data for the East Fork River, Wyoming. In a further study, Emmett (1976) reported that at lower values of stream power a river loses competence to transport the coarser bed particles and the channel bed becomes armoured. This limits the availability of smaller material.

The highest recorded rates for sediment transport in the Torlesse stream are shown in Fig. 10. It is assumed that these rates represent periods when sediment was freely available. These upper values tend to confirm Bagnold's view that, in this case, transport rates increase rapidly with increases in flow depth up to about 0.15 m. Thereafter, increases appear to be directly proportional to increases in flow depth. The upper limit values shown in Fig. 13 have been derived from a line fitted by eye through the upper values shown in Fig. 10.

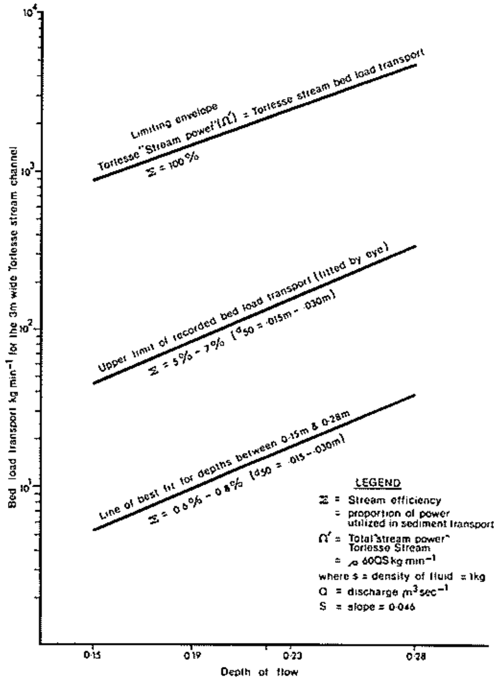

FIG.—13 Measured bed load transport rates and calculated stream power, Torlesse stream.

In Fig. 13 comparisons are made between 'stream power'* calculated for a range of flow depths for the Torlesse stream through the 3 m wide control section, the upper values of recorded bed load transport rates, and the mean values of transport rates within the depth range 0.15 m – 0.28 m. The median of gravel sizes ranged from 0.15 m – 0.30 m. The proportion of 'stream power' utilised in sediment transport (or efficiency Σ) when sediment was assumed to be freely available was found to be within the range of 5% – 7%. This finding is in remarkable agreement with the estimates of Leopold and Emmett (1976) who suggested efficiency values of 8% for d_{50}, 0.10 m and 5% for d_{50}, 0.50 m.

The proportion of 'stream power' expended in bed load transport in 'average' conditions (i.e. on the line of best fit) was found to be less than 1% (0.6% – 0.8%).

These estimates of stream efficiency have been made from data obtained over a limited range of flow depths and in consequence they should not be generalised. They do indicate however that stream power and stream efficiency can provide more reliable estimates of bed load, at least in this mountain stream, than can conventional estimating equations.

CONCLUSION

There are important differences in sediment relations between mountain and lowland streams, the most significant of which appears to be that bed load is not controlled by channel and hydraulic variables as is generally accepted in lowland streams. Therefore conventional bed load estimating equations are not appropriate for estimating bed load sediment yields. Although Bagnold's concepts of stream power and stream efficiency have been found very reliable they need further testing over a wider range of conditions before they can be recommended for general use.

The great variability in bed load sediment transport rates has been demonstrated. This, and the importance of low frequency events to total long term estimates of sediment yield, need to be taken into account if future studies are to yield reliable information. Despite this variability, results from the Torlesse stream catchment suggest that there is a *regime* condition of sediment yield which is related to peak flood discharge.

In the Torlesse stream catchment suspended sediments were found to contribute little to total sediment yield. As these results are contrary to some mountain catchment studies elsewhere in the world, further work needs to be carried out to determine the conditions under which suspended sediment may make a significant contribution to total sediment yield.

The results presented here have important implications for the management of mountain lands. They somewhat alter conventional views about the need and ability to manage certain mountain lands in order to minimise stream sediments.

ACKNOWLEDGEMENTS

I am particularly grateful to Messrs R. J. Blakely, P. Ackroyd and Le Ba Hong, for their invaluable assistance in the collection and processing of data presented here. Many people have assisted in weighing gravel at the sediment trap, usually

* Bagnold (1966) expressed available steam power as $\Omega = \rho, g, Q, s$.

Where ρ = density of fluid
g = gravity acceleration
Q = discharge
s = slope

When $\rho = 100\ \text{kg m}^{-3}$, $g = 9.81\ \text{m sec}^{-2}$, $Q = \text{m}^3\ \text{sec}^{-1}$; Ω = Newtons sec^{-1}.

This equation can be expressed as $\Omega' = \rho 60 Qs$ so that 'stream power' can be expressed in kg min^{-1} and therefore be comparable with measured rates of bed load transport (dry weight).

in wet and cold conditions, but my special thanks go to Messrs I. Fryer, R. A. Stratford and E. J. Costello. I am grateful to Mr G. W. Kitson for translating the paper by Ashida *et al.*

To Mr and Mrs Milliken, Broodsdale Station, my thanks for your hospitality and many kindnesses during the study period. I should also like to thank the Nuffield foundation for a grant of $3000.00 which allowed us to build the vortex tube sediment trap and base hut.

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SUSPENDED LOAD IN SOME MAJOR RIVERS OF NEW ZEALAND

S. M. Thompson and J. E. Adams

ABSTRACT

Suspended sediment data from two regions of New Zealand is used to illustrate an analysis of erosion. The large temporal and spatial variations of suspended load are described, and the consequent uncertainty in estimates of the mean load is indicated. The erosion, estimated as the mean load per unit of catchment area, is presumed to correlate with rainfall and the implied spatial variations in erosion are shown to agree with the erosion data. Thus a criterion for regionalisation is proposed which allows the erosion rate to vary within a region and requires only that the erosion response to a given rainfall is constant.

INTRODUCTION

In 1959 the M.W.D. established a programme of sediment gauging to estimate the erosion rate of upstream catchments, and provide background sediment data for the whole country. Nearly twenty years on, it is now time to review and analyse the accumulated data. Our present objective is to demonstrate how this can be done for one region and to provide a stimulus for similar work on other regions.

In providing for widespread and sustained gauging, the 1959 programme (Campbell, 1959) superseded occasional studies such as those of Marshall (1912), Benson (1946), and Benham (1948). Equipment and methods proven in the United States were introduced as described by Hopkins *et al.* (1959a,b,c) and Toebes (1963), and the results published in Ministry of Works *Hydrology Annual* numbers 5 to 15 between 1959 and 1970. This data reveals a wide range of erosion rates for the surveyed catchments. Within New Zealand's 260 000 km ^{2} land area there are catchments of 1000 km^2 eroding at up to 8000 t km^{-2}y^{-1}; (Jones and Howie, 1970) and others eroding at only 40 t km^{-2}y^{-1}; by world standards a wide range in a small area. This spatial variation, plus a more extreme and better appreciated temporal variation means that special analysis methods are required.

The objectives of this paper are to recognise parameters which are relatively invariant in time and space, and to describe a method of analysis which uses the parameters to estimate catchment erosion from imperfect data. As examples, the method is applied to hydro lakes Roxburgh and Matahina and to 33 catchments on the eastern slopes of the Southern Alps (Fig. 1). In discussing the eastern slopes of the Southern Alps we summarize all the available data and thereby indicate the scope of data now available for much of New Zealand. Our notation is defined where first used and also given at the end of the paper.

METHOD OF ANALYSIS

Sediment Ratings

Erosion rates are most frequently determined from suspended sediment data by the construction of sediment rating curves that relate instantaneous flow to sediment concentration. The method assumes that a rating curve, drawn to pass through a limited number of gaugings, can be used to convert a continuous time series of flow into a time series of sediment load. In this context each gauging is a measurement of sediment concentration based on laboratory analysis of a sample of the fluid, and a measurement of the flow based on a number of current

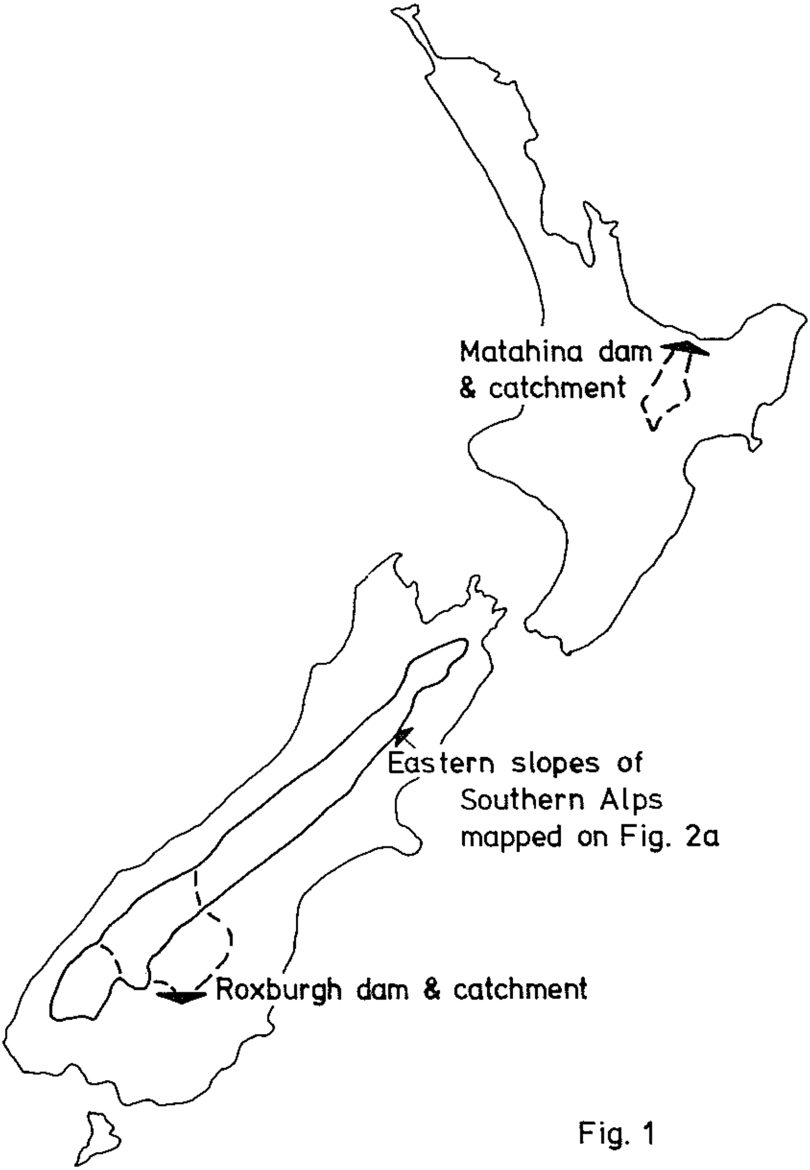

FIG.—1 Location map.

meter and depth measurements. The field measurements are done quickly and the result is assumed to describe the situation at one instant in time. Alternative methods of determining the sediment load involve daily or even continuous concentration determination, but require considerably more resources.

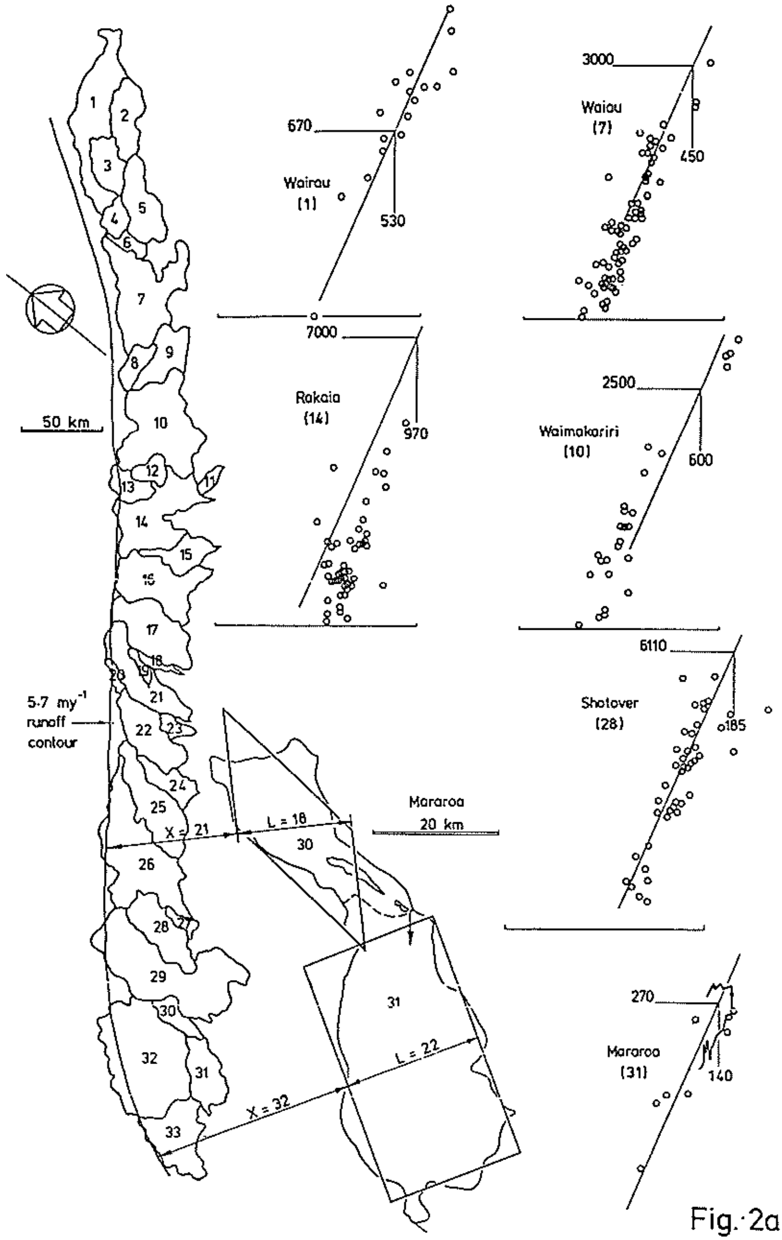

FIG.–2a and b: Log-log plots of concentration versus flow gaugings from 18 rivers and a map of 33 catchments on the eastern slopes of the Southern Alps, with an enlargement showing how the Mararoa is represented by two parallelograms.

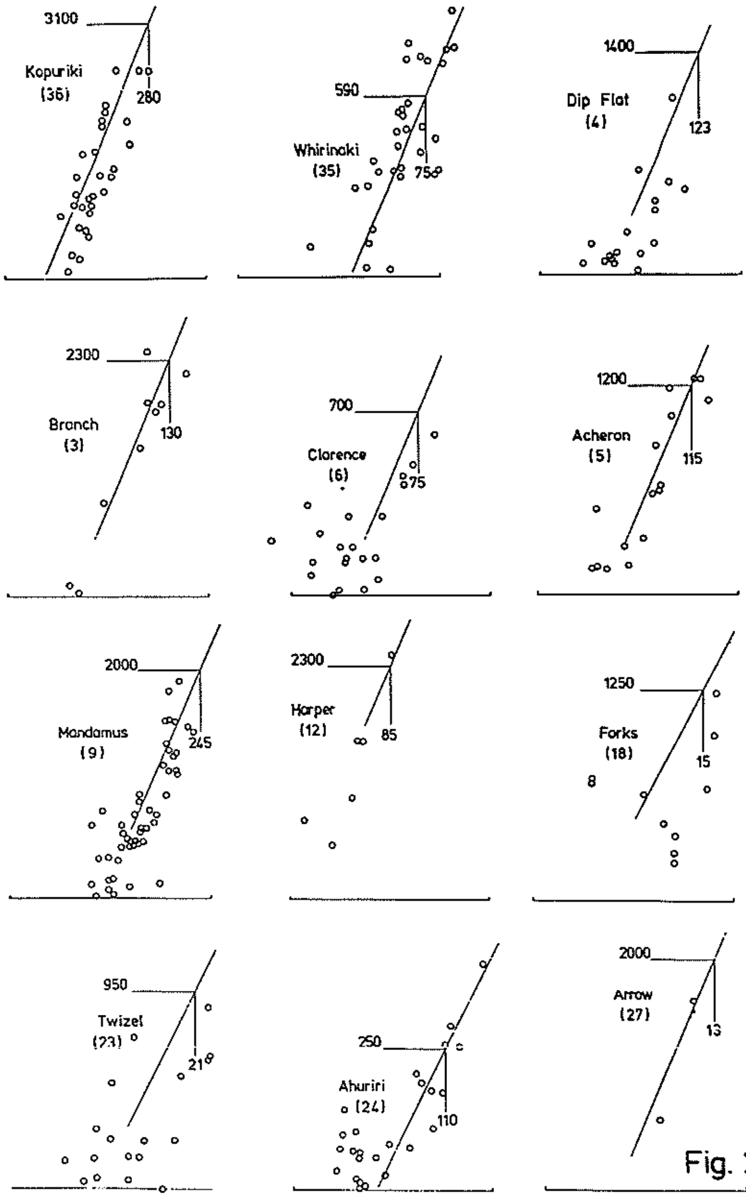

Fig. 2b

Figs. 2a and 2b give 18 log–log plots of concentration against flow, all but two of them from the eastern Southern Alps region. Lines at a 2.3 slope represent rating curves and the rated concentration at five times the mean flow is shown; for example at Wairau the concentration is 670 g m^{-3} at 530 m^{3}s^{-1}. The horizontal lines are two log units long and at the position of 10 g m^{-3}. In both Figs. 2a and

2b numbers in brackets below the catchment name refer to the map on Fig. 2a and to catchment numbers in Tables 1 and 4.

New Zealand rivers carry much lower suspended load concentrations than possible. In arid regions concentrations commonly reach $10^5 gm^{-3}$ but in New Zealand rarely exceed $10^4 gm^{-3}$ even in the largest floods. Hence the load carried must be controlled by the availability of load and not by capacity. In many floods the available load is exhausted early in the flood and the exhaustion causes a 'flushing effect' with more load carried during rising than during falling river flow.

Within New Zealand direct knowledge of the flushing effect is almost entirely restricted to data from the Mararoa River in Southland. However the flushing effect shows indirectly in data from almost all rivers. When concentration of suspended load is plotted against flow (generally on log–log paper) the scatter of the points about the best–fit line is much greater than can be explained by errors in sampling either concentration or flow, and is roughly what would be expected from the flushing effect. The flushing effect causes a 'loop rating' of concentration to flow in contrast to the straight line relationship commonly assumed.

New Zealand rivers rise and fall rapidly, and because most rivers are distant from where the samplers live most measurements have been made during the falling stage of floods. In order to estimate suspended load straight lines are fitted by eye, and drawn midway between the scatter of points and not through the centre of gravity of the points. In other words because many of the points are on the falling stage and few are on the rising stage, the lines lie above most of the points and above the least squares regression line. As might be expected all lines that allow for the flushing effect give larger estimates of the suspended load than do regression lines giving equal weight to all points.

A New Zealand flood that was measured in unusual detail to obtain data for design of an operating rule for the Lake Manapouri control gate, occurred on the Mararoa River (46°S, 176°E) on 3 May, 1976 (Christian and Thompson, 1978). The sediment concentration during the rising stage greatly exceeded the concentration at the same flow during the falling stage, so that the time plot of concentration versus flow is a loop (Fig. 2a). The scatter of gaugings about the rating line is similar at many other sites, and is also believed due to loop rating behaviour, although there is very little published New Zealand data from large catchments supporting this belief.

Loop rating behaviour of the same type has been documented by detailed measurements made by Wood (1977) from a 154 km² English catchment and by Schouten (1977) from a 2.5 km² New Zealand catchment. Apparently the width of the loop, the difference between rising and falling stage sediment concentration, is less for the larger catchments such as the Mararoa. Wood shows differences in the inclination of the loop axis between floods and suggests why these occur, but we have no New Zealand data to test his hypothesis.

We will represent the rating curve by the formula:

$$C_i = C_5\,(q/5)^e \qquad (1)$$

which is a straight line on log–log graph paper and so can be easily fitted to scattered data by eye. C_i and C_5 are the instantaneous concentration and the concentration corresponding to five times the mean flow given by the rating curve. C_5 may be seen as a pivot point o a set of rating curves which plot as straight lines on a log–log plot with different slopes, e. Alternatively C_5 may be seen as a coefficient that is not dependent on the slope e. Typically $400 < C_5 < 4000$ g m^{-3} in New Zealand. Q_i, Q and q are respectively the instantanteous flow, mean flow and the ratio Q_i/Q.

Flow Duration Curve

A flow duration curve is a graph of the fraction of time f that the flow exceeds q plotted against q. The theory of continuous variables and their extremes predicts that, in the extreme, f decreases exponentially:

$$f = \exp(-aq - b), \tag{2}$$

where a and b are constants. Thus a graph of the logarithm of f against q tends to be a straight line at the extreme. The measured flow duration curves from New Zealand rivers (Fig. 3 and Table 1) do exhibit this form despite short flow records and probable errors in some stage to flow conversions.

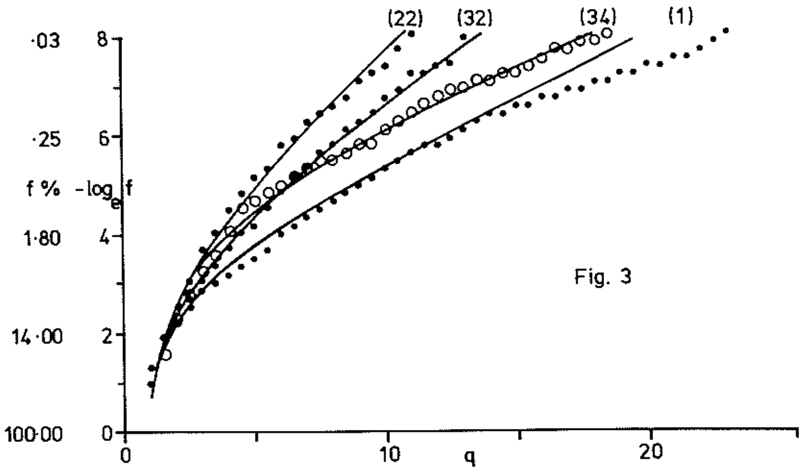

FIG.–3: A log-natural plot of flow duration curves for: (22) Lake Ohau inflow a = 0.50, b = 3; (32) Lake Te Anau inflow 0.38, 3; (34) Manuherikia 0.24, 4; (1) Wairau at Tuamarina 0.26, 3. The dots represent measurements and the lines represent equation 3.

When flow data is plotted as $-\log_e f$, ordinate, against q, abscissa, (Fig. 3) then 'a' is the slope of the asymptote at large q, and 'b' is the ordinate of this asymptote at $q = 0$. Larger values of 'a' describe either records from catchments with more flood storage and consequently slow recession rates from flood peaks, or records in which a large part of the annual water yield is spring snow melt. Larger values of 'b' describe records from catchments where more rainfall is absorbed before any runs off and the absorbed water is later evaporated and does not run off.

Equation (2) for the asymptote to the flow duration curve is fitted to lower flows by adding the expression $(a + b - 0.7)/q$ as follows:

$$f = \exp(-aq - b + (a + b - 0.7)/q) \tag{3}$$

Thus the curve passes through $f = 0.5$ at $q = 1$, as if the mean flow equals the median flow. In practice the mean always exceeds the median and so equation (3) does not fit flows near the mean. Although substitution of 1.1 for 0.7 in equation 3 does allow the equation to fit flows near the mean, it does not affect estimates of the long term sediment load by more than 1% and does not fit the flow data as well at the critical higher flows. These details of curve fitting are not important in the context of sediment load and so have not been pursued here.

TABLE—1: Flow duration curves. Entries in this table are the fraction of time that the flow exceeds q times its mean expressed as %, for sites with good data. Italics represent equation 2 for intermediate values of a, and the other figures are measured values.

	Catchment	*a*	*q* = 2	*4*	*8*	*12*	*16*	*years of record*	*end of record*	
	b = 3									
		0.20	*12*	*4.2*	*1.4*	*0.56*	*0.24*			
2	Waihopai	0.23	10	3.8	1.1	0.42	0.18	16.1	Aug	76
		0.25	*11*	*3.5*	*0.93*	*0.31*	*0.11*			
1	Wairau-Tuamarina	0.26	10	3.9	0.91	0.28	0.11	15.0	Jun	75
		0.32	9.7	2.6	0.53	0.13	0.03			
23	Twizel	0.32	10	2.4	0.47	0.09	0.04	7.9	Jun	70
26	L. Wanaka inflow	0.32	8.4	2.2	0.51	0.13	0.04	46.5	Jun	76
5	Acheron	0.36	9.7	2.1	0.33	0.08	0.05	18.6	Dec	76
32	L.Te Anau inflow	0.38	10	2.4	0.28	0.06	0.02	49.6	Sep	75
25	L.Hawea inflow	0.38	9.5	2.0	0.31	0.06		46.1	Jan	76
31	Mararoa	0.40	10	2.2	0.12	0.02		2.9	Jly	77
		0.40	*8.6*	*2.0*	*0.28*	*0.05*	*0.01*			
28	Shotover	0.42	9.0	1.6	0.22	0.05		9.1	Aug	76
9	Hurunui-Mandumus	0.44	7.4	1.4	0.23	0.05	0.01	20.4	Mar	77
27	Arrow	0.46	10	1.2	0.18			2.2	May	76
		0.50	*7.4*	*1.4*	*0.13*					
22	L.Ohau inflow	0.50	7.7	1.1	0.13	0.02		50.8	Jly	77
19	Jollie	0.54	7.5	0.99	0.09	0.01		10.7	Dec	76
18	Forks	0.54	7.4	0.84	0.08	0.01		14.4	Jan	77
		0.62	*6.2*	*0.87*	*0.05*					
	b = 4									
3	Branch	0.14	10	2.4	0.77	0.37	0.22	9.7	Apr	76
		0.16	*7.5*	*2.3*	*0.78*	*0.36*	*0.18*			
11	Selwyn	0.17	9.0	2.4	0.75	0.26	0.13	13.0	May	77
		0.20	*7.1*	*2.0*	*0.57*	*0.22*	*0.09*			
–	Pomahaka	0.20	10	2.5	0.55	0.16	0.07	15.1	Sep	76
34	Manuherikia-Ophir	0.24	10	1.6	0.40	0.10	0.05	4.8	Dec	76
		0.25	*6.6*	*1.6*	*0.39*	*0.12*	*0.04*			
6	Clarence	0.28	9.5	1.7	0.31	0.08	0.03	18.6	Dec	76
7	Waiau-Marble Pt.	0.28	8.0	1.7	0.32	0.08	0.02	9.1	May	77
29	L.Wakatipu inflow	0.30	8.3	1.5	0.20	0.06	0.02	50.3	Jly	76
		0.32	*5.9*	*1.3*	*0.22*	*0.05*	*0.01*			
24	Ahuriri	0.32	7.7	1.2	0.23	0.04	0.01	11.8	Jun	25
4	Wairau-Dip flat	0.34	7.5	1.2	0.17	0.03		25.7	Jan	77
		0.40	*5.2*	*0.93*	*0.12*	*0.02*				
	b = 5									
		0.25	*3.4*	*0.77*	*0.16*	*0.05*	*0.02*			
35	Whirinaki	0.25	7.9	0.90	0.11	0.04	0.02	24.4	Apr	77
37	Rangitaiki-Kopuriki	0.57	2.6	0.23	0.05	0.01		10.9	Jun	77
		0.57	*2.5*	*0.23*	*0.05*	*0.01*				

Mean Suspended Load

We may calculate the mean load as the sum of the products of each value of flow by the concentration at that flow and the fraction of time that the flow has that value:

$$G = \sum_{f=0}^{h} Q_i \, C_i \, df \tag{4}$$

However, now that we have assumed explicit algebraic formulas for the sediment rating (Eqn. 1) and flow duration curve (Eqn. 3), we may derive a more convenient formula for the load:

$$G = h \, C_5 \, Q \tag{5}$$

The dimensionless coefficient h is the integral:

$$h\,(a, b, e) = \int_{q\min}^{\infty} (C_i/C_5) q(-df/dq)\, dq$$

$$= \int_{q\min}^{\infty} 5^e (aq^{e+1} + (a+b-0.7)q^{e-1}) \exp(-aq - b + (a+b-0.7)/q) dq$$

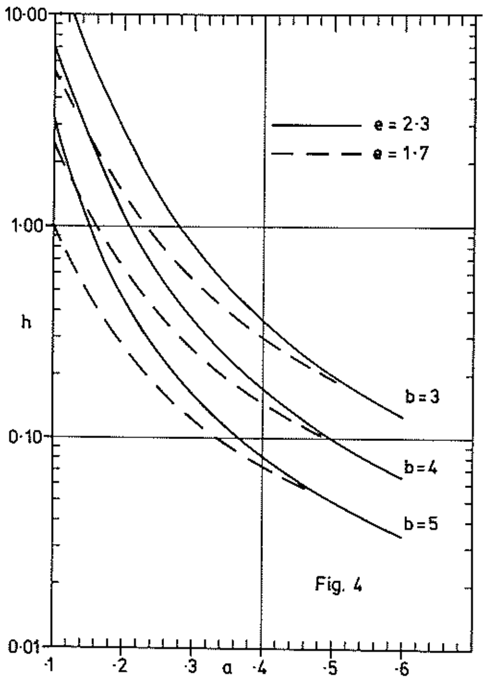

FIG.–4: A log-natural plot of the function h defined below equation 5 in terms of *a*, *b* and *e*, which represents the ratio of the mean transport to the product of the mean flow and the concentration C_5.

and has been evaluated numerically for some useful values of a, b and e and graphed in Fig. 4.

To estimate suspended load by equation 5, a log–log plot of concentration against flow is required to determine C_5 and e (Fig. 2), a flow duration curve is required to determine Q, a and b (Fig. 3) and Fig. 4 is used to find h.

For the measured range of a, b and e the integrand varies with flow q in the manner shown in Fig. 5. The integrand represents the fraction of load per unit range of flow that occurs at each flow and has a single maximum. For example when $a = 0.32$ and $e = 2.3$, the maximum is at $q = 8$ because although larger flows carry enormous loads they are also very rare while the far more frequent, smaller flows carry only small loads. Although flows less than some small value q_{min} contribute a negligible fraction of the total load, the assumed formula for the flow duration curve, in association with the sediment rating formula, gives a spurious second maximum of the integrand at these small flows. To resolve this difficulty the numerical integration to obtain h is performed only for flows greater than $q = 2.5$, and 1.5 times integrand at $q = 2.5$, is added to allow for loads at smaller flows. The justification for this arbitary procedure is the corresponding plausible shape of the integrand curves in Fig. 5.

Fig. 4 provides a sensitivity test to estimate error bounds on load from the uncertainties in the values of a, b, and e and hence h. Measurement of parameter 'a' depends critically on the accuracy of the stage to flow rating at high flows. It also depends on there being sufficient length of record to include an adequate sample of high flows. This second requirement can be quantified by the following analysis of the long Lake Wanaka inflow record. Flood flows estimated from the rate of change of lake level are more reliable than estimates using a stage to flow conversion that is an extrapolation from river gaugings at lower flows. The Wanaka record is described by $a = 0.32$ and $b = 3$; and from equation 3 at $q =$

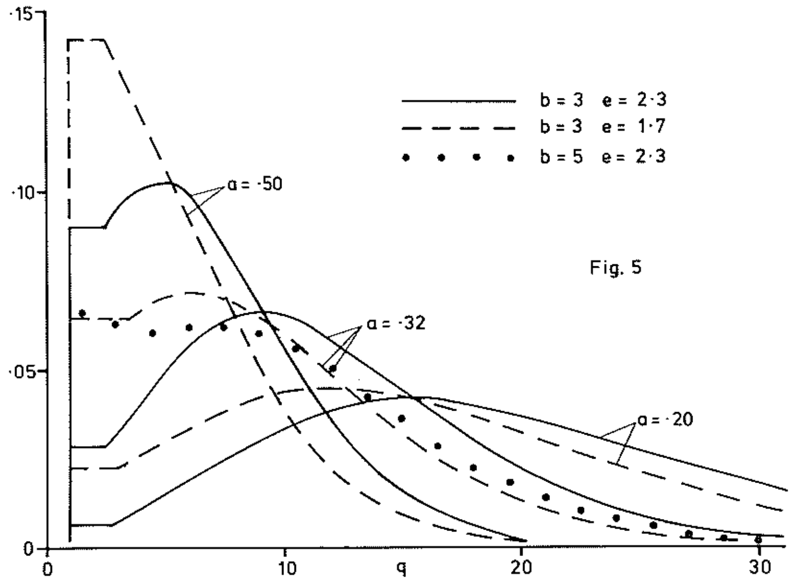

FIG.–5: A plot of the integrand in the equation which defines h. It shows for instance that flows above 16 times the mean contribute about 2%, 20% and 50% of the mean transport when a = 0.50, 0.32 and 0.20 respectively. The ordinate scale is set so that there is unit area under each curve.

8, $da/df = -0.25$; and for $e = 2.3$, $h = 0.67$ and $dh/da = -6$ from Fig. 4. so to estimate h within 50% with 95% confidence we require an expected standard deviation on $f(8)$ less than $0.50 \times 0.67/1.96 \times 0.25 \times 6 = 0.11$. By dividing the Wanaka record into many short records we obtain the estimates for the standard deviation of $f(8)$ in Table 2. These estimates of standard deviation become unreliable for the longer record lengths because the samples are few, and so the estimates fluctuate about the true value which can be expected to decrease monotonically. Table 2 is extended to 9 year records so that this fluctuation can be seen and averaged out, and from the table we are able to conclude that 5 years of record are necessary to estimate the load to within 50%.

TABLE—2: The standard deviation of the fraction of time that Lake Wanaka inflow exceeds 8 times its mean.

Record length in years	1	2	3	4	5	6	7	8	9
Number of records	36	18	12	9	7	6	5	4	4
Standard deviation of $f(8)$	0.27	0.16	0.15	0.12	0.09	0.11	0.06	0.12	0.08

Sensitivity analyses of other parameters will provide similarly quantitative results. Thus the way is opened by this method of analysis to establish regional values of the parameters for application to less well measured catchments.

APPLICATIONS

Siltation of Hydro Lakes

Roxburgh Dam (45°S, 169°E) and Matahina Dam (38°S, 177°E) have formed small lakes which have trapped all the bedload and about two thirds of the suspended sediment in their respective rivers. Regular bathymetric surveys of the lakes are undertaken because siltation will eventually hinder operation of the associated power stations. These surveys provide an independent estimate of the suspended sediment transport (Table 3) which agrees with the gauged erosion rates of tributary catchments. The estimates of density, bedload and passing fraction in Table 3 are in the ranges of values suggested in the 'ASCE Sedimentation Manual' (Vanoni, 1975).

Gaugings of upstream tributaries provide estimates of erosion rates for the catchments of Roxburgh and Matahina (Table 3). There is wide variation within each catchment. The Matahina catchment comprises rapidly eroding upthrust greywacke ranges and slowly eroding permeable pumice lowlands. The Roxburgh catchment comprises a rapidly eroding wet region and a slowly eroding dry region. Clearly the large spatial variations in erosion rate due to geology and rainfall must be taken into account when siltation is predicted from gaugings of upstream tributaries.

Erosion of the Eastern Slopes of the Southern Alps

The following analysis shows how the spatial variation of erosion on the eastern slopes of the Southern Alps is consistent with a relationship between rainfall and erosion.

Coulter (1973) presented a map of New Zealand runoff showing that runoff decreases 'exponentially' south–eastwards from the main divide of the Southern Alps. For the present work it was determined by fitting representative raingauge and catchment runoff data that runoff halves for each 12.5 km away from the divide.

$$R_x \;\alpha\; \exp(-x/18). \tag{6a}$$

where R_x is the runoff in m y^{-1}, x km from the divide.

The mean annual runoff of catchments draining the eastern slopes of the Southern Alps is thus related to their shape and position with respect to the divide. If L km is the NW-SE length of the catchment and the catchment runoff, R m y^{-1} is known, then:

$$X = 18 \log_e \left[\frac{5.7}{R} \left(\frac{1-\exp(-L/18)}{L/18} \right) \right] \tag{6b}$$

is the distances (in km) from the headwaters of the catchment to the 5.7 m y^{-1} runoff contour.

TABLE–3: Siltation data from hydro lakes Roxburgh and Matahina, compared with gauged erosion data from Table 4. Matahina siltation data from Callander and Duder (1977).

Siltation Data	*Unit*	*Roxburgh*	*Matahina*
Period between bathymetric surveys		1961-1974	1966-1975
Volume of deposit	hm^3	18.8	1.63
Density of deposit	t m^{-3}	1.27	1.05
Mass of deposit	Mt	23.9	1.71
Total deposition	Mt y^{-1}	1.84	0.19
Bedload	Mt y^{-1}	0.18	0.04
Hence suspended sediment deposition	Mt y^{-1}	1.66	0.15
Suspended sediment passing dam	Mt y^{-1}	0.93	0.08
Total suspended sediment input	Mt y^{-1}	2.59	0.23

	Gauged catchment erosion rates	*Land Area km²*	*Erosion t km⁻² y⁻¹*	*Yield Mt y⁻¹*
28	Shotover	1082	2040	2.21
27	Arrow	196	200	0.04
34	Rest like Manuherikia	7584	45	0.34
	Total to Roxburgh	8826	293	2.59
35	Whirinaki	530	130	0.07
	Murupara	1180	31	0.04
	Remainder to Kopuriki	610	166	0.10
36	Total to Kopuriki	2320	90	0.21
	Remainder to Matahina	380	50	.02
	total to Matahina	2700	85	.23

TABLE—4: Gauged erosion data from the eastern slopes of the Southern Alps, and from the catchment of Matahina Dam. Numbered catchments are mapped on Figs. 1 and 2a. The parameters and equations relating them are defined in the text.

	Catchments Mapped on Figure 2a	A km^2	Q $m^3 s^{-1}$	R $m\,y^{-1}$	L km	X km	a	b	h	C_5 $g\,m^{-3}$	Y $Mt\,y^{-1}$	E $t\,km^{-2}\,y^{-1}$	E_o $kt\,km^{-2}\,y^{-1}$
1	Wairau-Tuamarina	3430	121	1.11	52	9	0.26	3	1.2	670	3.07	895	17
2	Waihopai	764	17	0.70	17	30	(0.23)	3	1.8	1200	1.16	1510	(132)
3	Branch	580	26	1.41	20	16	(0.14)	4	2.7	2300	5.07	8750	(160)
4	Wairau-Dip flat	505	25	1.56	18	15	0.34	4	0.26	1400	0.29	570	9
5	Acheron	997	23	0.73	28	25	0.36	3	0.47	1200	0.41	410	29
6	Clarence	448	15	1.06	33	16	0.28	4	0.43	700	0.14	320	9
7	Waiau-Marble Point	1980	90	1.43	50	5	0.28	4	0.43	3000	3.66	1850	21
8	Lake Sumner	342	25	2.31	20	7							
	rest to Mandamus	730	24	1.04	35	16	—	—	—	—	0.84	1140	33
9	Hurunui-Mandamus	1072	49	1.44	38	9	0.44	3	0.27	2000	0.84	—	—
10	Waimakariri-Gorge	2460	120	1.54	52	3	0.38	3	0.40	2500	3.79	1540	14
11	Selwyn	150	3.1	0.65	13	33	0.17	4	1.8	420	0.07	490	52
12	Harper	320	17	1.68	20	13	0.30	4	0.35	2300	0.43	1350	17
13	Wilberforce	480	47	3.09	30	−2	0.35	3	0.52	2400	1.85	3850	11
14	Rakaia-Gorge	2580	194	2.37	43	−2	0.50	3	0.20	7000	8.56	3320	14
15	Ashburton-Mt Somers	610	9	0.47	40	28	0.40	3	0.35	650	0.07	110	16
16	Rangitata-Gorge	1494	92	1.94	45	1	0.50	3	0.20	4500	2.61	1750	11
17	Lake Tekapo	1425	75	1.66	44	4	0.37	3					
18	Forks	130	2.9	0.70	26	26	0.54	3	0.16	1250	0.02	140	11
19	Jollie	139	7.1	1.61	12	17	0.54	3	0.16	(550)	0.02	140	(2)
20	Hooker	122	21	5.43	8	−3	0.33	3	0.60	1600	0.64	5210	(6)
21	Lake Pukaki	1360	118	2.74	50	−6	0.45	3					
22	Lake Ohau	1190	80	2.12	41	1	0.50	3					

23	Twizel	261	4.2	0.51	20	34	0.32	3	0.67	950	0.08	320	51
24	Ahuriri	566	22	1.23	35	13	0.32	4	0.30	(250)	0.06	90	(2)
25	Lake Hawea	1470	63	1.35	42	9	0.38	3					
26	Lake Wanaka	2545	189	2.34	49	–3	0.32	3					
27	Arrow	196	2.5	0.39	18	40	0.46	3	0.23	2000	0.04	200	55
28	Shotover	1082	37	1.10	38	14	0.42	3	0.31	6110	2.21	2040	54
29	Lake Wakatipu	2980	156	1.66	80	–5	0.30	4					
30	South Mavora Lake	346	12	1.10	18	21							
	rest of Mararoa	873	16	0.57	22	32	—	—	—	—	0.09	95	13
31	Mararoa	1219	28	0.72	—	—	0.40	3	0.35	270	0.09	—	—
32	Lake Te Anau	3355	263	2.48	46	–3	0.38	3					
33	Lake Manapouri	1270	132	3.28	30	–3							

Catchments in Table 3

34	Manuherikia-Ophir	2144	15	0.22	—	—	0.24	4	0.52	400	0.09	45	—
35	Whirinaki	534	15	0.89	—	—	0.25	5	0.25	590	0.07	130	—
36	Rangitaiki-Kopuriki	2318	56	0.76	—	—	0.57	5	0.038	3100	0.21	90	—

In the case of the Mararoa (Fig. 2a) the catchment has a lake in the middle and two parallelograms must be used: an upstream one ($L = 18$, $X = 21$) that includes the lake and a downstream one ($L = 22, X = 32$). Because of the lake the upstream parallelogram provides water but no sediment and the downstream one both water and sediment.

Calculated values of X (Table 4) define the position of the 5.7 m y^{-} runoff contour with a probable error of 4 km which is equivalent to a 20% error in runoff. According to the exponential relationships adopted here eastern runoff and precipitation are determined by the position of the 5.7 m y^{-1} runoff contour and an identical 6400 mm y^{-1} precipitation contour. The contour is roughly parallel to the divide but lies to the west in the north and to the east of the divide in the south (Fig. 2a). The Southern Alps with its long narrow strip with very high rainfall is thus somewhat unusual in its rainfall regularity.

The runoff–distance relationship, equation (6a), can be converted to a precipitation–distance relationship by allowing for evaporation, and the precipitation P_x in mm y^{-1}, x km from the divide is approximately

$$P_x = 6400 \exp(-x/22) \tag{6c}$$

Fournier (1960) proposed a universal formula for erosion which is based on measurements from all parts of the world, including the schist mountains of the Rhone valley which are similar to the Southern Alps. The formula is:

$$\log_{10} E = 2.65 \log_{10} (p^2 P) + 0.46 \log_{10} (H^2/A) - 1.56 \tag{7a}$$

Where E = erosion in t km^{-2} y^{-1}, p = precipitation in the wettest month as a fraction of P, H = mean catchment elevation above outlet in m, A = catchment area in km^2. In the Southern Alps the seasonal pattern of rainfall varies very little and may be assumed constant ($p \approx 0.1$). Also the catchments have similar topography and the value of H^2/A may be assumed constant. Then

$$E_x \propto P_x^{2.65} \tag{7b}$$

and combining with equation 6c:

$$E_x = E_o \exp(-x/8.3) \tag{8a}$$

where E_o is the erosion for 5.7 m of annual runoff. Thus the average erosion for an eastern slope catchment in terms of X and L will be

$$E = \frac{E_o}{L/8.3} \left[\exp(-X/8.3) - \exp(-(X+L)/8.3) \right] \tag{8b}$$

or $$E_o = E \exp(X/8.3) \left[\frac{L/8.3}{1-\exp(-L/8.3)} \right] \tag{8c}$$

Estimates of the average erosion E using equation (5), for those catchments on the eastern slopes of the Southern Alps that have sufficient flow and sediment data are presented in Table 4. Regional concentration-flow data was examined and it justifies our choice of $e = 2.3$ throughout. These data and equation (8c) allow E_o to be estimated from each catchment. For most large catchments with reliable data the values of E_o are in good agreement. The average value weighted by catchment area, is about 18 000 t km^{-2} y^{-1} and might be used to estimate erosion of ungauged catchments.

Three small catchments in the Waitaki basin give low estimates of E_o. For the Jollie and Ahuriri catchments the lack of scatter in the concentration-flow plots suggests that all measurements were made after the flood peak and therefore according to the flushing effect give spuriously low loads. The Hooker catchment is mostly glacial and presumably requires special analysis which has not been attempted. Two catchments in the Wairau basin give high estimates of E_o. Their load exceeds that measured downstream and is associated with abnormally large values of h in turn due to abnormally small values of a. Overestimated flows during floods would cause load estimates like these. The high but consistent values of E_o for the Arrow and Shotover relative to the rivers to the north can be explained in terms of lithology – schist versus greywacke. Schist is softer and weaker than greywacke and so can be eroded more easily. The values of E_o suggest that schist is eroded three times faster than greywacke for the same precipitation.

CONCLUSION

Our analysis indicates that 5 years of excellent flow records and a minimum of 2 sediment gaugings at flows about 5 times the mean, one on a rising and one on a falling stage, are necessary for useful prediction of long term sediment load. At many sites the gaugings to date have not provided this minimum data. Sediment sampling at flows less than the mean does not help estimate long term sediment load because the low flows transport such a small part of the load. The time of each gauging must be recorded and subsequently plotted on the hydrograph to establish the position and magnitude of the associated flood peak. This information relating to the flushing effect must then be used with the plot of concentration versus flow when the rating curve is drawn.

We have proposed parameters (in particular $a,b,e; h,C_5$) which describe the essential features of the flow duration curve and suspended sediment transport, and which seem to have constant values within suitably defined regions. Evidence is required to confirm the regional homogeneity, but ultimately it is hoped that regional values for parameters like these will be established and will make routine field gauging unnecessary. Field gauging will then only be required for specific important projects to confirm estimates based on established regional statistics.

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Notation

Dimensionless parameters:

a, b	describe flow duration curve
e	exponent of sediment rating
f	fraction of time flow is exceeded
h	ratio G/C_5Q
p	precipitation in wettest month as a fraction of P
q	ratio Q_i/Q

Dimensioned parameters describing *long term mean* properties of a whole catchment:

A	km^2	land area
C_5	$g\ m^{-3}$	sediment concentration at five times mean flow
E	$t\ km^{-2}y^{-1}$	erosion $= Y/A = 31.56\ G/A$
G	$g\ s^{-1}$	transport of suspended sediment
H	m	height of mean elevation above outlet
L	km	length normal to divide
P	$mm\ y^{-1}$	precipitation
Q	$m^3\ s^{-1}$	flow
R	$m\ y^{-1}$	runoff $= 31.56\ Q/A$
X	km	distance of headwater from $5.7 m\ y^{-1}$ runoff contour
Y	$t\ y^{-1}$	suspended sediment yield $= 31.56\ G$

Dimensioned parameters describing instantaneous or local properties:

C_i, Q_i	instantaneous values of concentration and flow
E_x, P_x, R_x	local erosion, precipitation and runoff x km from $5.7\ m\ y^{-1}$ runoff contour
E_o	local erosion at $5.7\ m\ y^{-1}$ runoff contour
x	distance from $5.7\ m\ y^{-1}$ runoff contour

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