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LATE CENOZOIC EROSION IN NEW ZEALAND

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ABSTRACT

Uplift and erosion are roughly equal in the Southern Alps of New Zealand and the following rates have been determined:

tectonic uplift	$620 \pm 20 \text{ Mt y}^{-1}$
river load	$700 \pm 200 \text{ Mt y}^{-1}$
offshore deposition	$580 \pm 110 \text{ Mt y}^{-1}$

The tectonic uplift is the result of oblique collision between the Indian and Pacific plates, with the edge of the Pacific plate being upturned and uplifted as the Southern Alps, crustal narrowing of 22 mm y^{-1} being converted to uplift along a curved fault plane. (p. 14)

Almost all rock eroded from the Southern Alps is carried as suspended load by rivers. River bedload is of minor importance, and its abrasion adds to the suspended load. The estimated suspended load amounts to 265 Mt y^{-1} , but with a single exception only normal loads have been sampled, and the additional abnormal load from earthquake-caused landslips is estimated to double the normal load. The river load estimate is confirmed in part by spot checks from sediment accumulated in onshore traps.

A model proposed for the growth of the Southern Alps from a peneplain shows that the range attained steady state about 1.5 My after uplift started. With uplift initial non steady state, flat topped mountains like those that remain in Otago, become steady state spiky mountains. The range as a whole is in steady state, though the individual mountains change. The offshore deposition rates agree with the river load and tectonic uplift estimates and thus provide substantial confirmation for the steady state model.

check
w/ Albert

In the South Island of New Zealand the pattern of uplift is simple and the uplift and erosion rates are so high that man accelerated erosion is a negligible part. By contrast, in the North Island the pattern is less simple, the rates lower, and man accelerated erosion more significant. All available information relevant to the last two million years of uplift and erosion in the South Island has been examined. The viewpoint taken is from geology ($>10^3$ y) and not from engineering or culture ($<10^2$ y).

None of the information in the thesis is particularly accurate, but none can be disregarded, and so assumptions are made to assemble a coherent and consistent whole. In doing this considerable use is made of exponential formulae similar to those used by seismologists in their earthquake magnitude scale.

In New Zealand early work on erosion rates was sporadic and done mainly by engineers. Cotton and other geomorphologists of the Davisian school considered that landforms took tens of millions of years to form. They assumed ample time was available and were not interested in rates. Furkert, an engineer concerned with harbour protection, made the first reasonable estimates.

This thesis was started in late 1974 and finished in early 1978. It is only within the last few years that the necessary regional data has become available. For the first time uplift rates had been quantified, land retriangulations analysed, river flow records computerized, and the country mapped topographically at 1:63 360, and mapped geologically at 1:250 000. Hydro lake siltation problems are presently providing considerable economic impetus for river load studies.

Lack of a good name for the subject of this thesis (broadly: uplift, erosion, and mountain growth) lead to problems in literature review, and probably not all important papers have been found. The total number found is about ten times that cited as references. Four kinds of data were used: rates taken directly from published work; rates inferred from published work related to, but not interpreted in terms of, uplift and erosion; rates from published (e.g. topographic maps) and unpublished (e.g. computer files) data not directly related to uplift and erosion; and data from my own field observations during 20 weeks in the South Island (except Fiordland) and from my experiments. The approximate ratio of each kind is 1:10:80:10. Much of the data is unpublished, and government and local body departments were

approached directly with requests for specific information not yet published.

Erosion rates taken from the literature are given in appendix 1 for 16 parts of the South Island. The methods used to determine the rates vary too much in nature and reliability for the rates to be used in the regional analysis, but the rates themselves are consistent with those in the text.

All figures and tables are placed at the end of the chapter in which they are first referenced to avoid disrupting the text.

Units are S.I. where possible, and quoted imperial values have been converted to S.I. at comparable accuracy. The following non-standard units are used: $t=10^3$ kg, $kt=10^6$ kg, $Mt=10^9$ kg, $Gt=10^{12}$ kg; $y=31.56 \times 10^6$ s, $My=10^6$ y; $ppm=10^{-3}$ kg m^{-3} . Where not stated, standard densities of 1.25 (silts), 1.8 (silty gravel), 2.1 (gravel), and 2.65 (bedrock) $t\ m^{-3}$ have been used to convert volumes to masses.

Notation used only locally in the thesis is defined where used, but notation used consistently in many parts of the text and in tables is defined where first used and given below for reference.

Notation	units	use
A	km^2	catchment area
C	ppm	suspended load concentration
E	$t\ km^{-2}\ y^{-1}$	annual erosion
G	$10^{-3}\ kg\ s^{-1}$	river load transport rate
P	$mm\ y^{-1}$	annual precipitation
Q	$m^3\ s^{-1}$	river flow
R	$m\ y^{-1}$	annual runoff
S	rad.	channel slope
Y	$Mt\ y^{-1}$	annual load

An annotated bibliography of otherwise unreferenced sources that include maps, photographs and references for readers unfamiliar with New Zealand geography, topography and geology follows.

Cotton, C. A. 1960: "Geomorphology, an introduction to the study of landforms" 7th rev. ed. Whitcombe and Tombs, Christchurch, New Zealand. 505 pp. Classic cycle approach to landforms illustrated mainly with New Zealand examples.

Collins, B. W.; Fraser, R. 1971: "Recent crustal movements" Royal Society of New Zealand Bulletin 9:247 pp. About half the papers are on New Zealand topics, which include the Inangahua earthquake, seismicity, and coastal deformation.

Department of Scientific and Industrial Research: "Geological map of New Zealand 1:250 000". Wellington, New Zealand. Coverage of entire country on 27 sheets with yard grid, and spot heights. Sediments mapped by age, and many active and inactive faults shown.

Kingma, J. T. 1974: "The geological structure of New Zealand" Wiley, New York, 407 pp. Largely written pre-plate tectonics but contains excellent oblique air photographs.

Lands and Survey: New Zealand Map Series 1. Topographic maps covering the South Island on 184 sheets at 1:63 360 scale with a 1000 yard grid and 100 or 250 foot contours. Published c. 1943 to present, most post-1960 maps compiled photogrammetrically.

MacKenzie, I. (Ed.): New Zealand Journal of Geology and Geophysics. Department of Scientific and Industrial Research, Wellington. Published bimonthly, and amounting to about 1000 pages per year. Main publication for New Zealand geology.

Wards, I. (Ed.) 1976: New Zealand Atlas. Government Printer, Wellington, New Zealand. 292 pp. Maps, articles and photographs of New Zealand. Regional maps at 1:1 000 000 show most localities mentioned in the text, and the excellent photographs show many of the landforms discussed.

CHAPTER TWO

TECTONIC DEFORMATION AND UPLIFT
IN NEW ZEALAND

There are at least four different ways of estimating the nature and rate of tectonic deformation for New Zealand, and all give a rate that is extremely high by comparison with most other places in the world. Plate tectonics is the most basic. In plate tectonics a finite number of rigid plates are assumed to cover the earth. The three of particular importance for New Zealand are the Indian (which includes Australia), the Pacific, and the Antarctic. New Zealand, other than the Northland Peninsula, lies along the Indian-Pacific plate boundary (Fig. 1), and so it is important to know the nature of the relative movement between these two plates and the rate of movement. Although the plates are assumed rigid, within New Zealand the plate boundary is a belt and not a sharp line. The belt is about 60 km or so wide in the middle of the South Island and up to 150 km wide elsewhere.

Because the plates are assumed rigid and the earth is round there are always a pair of points that define the axis around which the relative movement of a pair of plates takes place. It is sufficient to define one of the points (the other being antipodal) and the one chosen here, not unnaturally, is the nearest. It is a pole of rotation, but must be distinguished from the geographic rotational poles, and is called a 'shift-pole' here. By necessity the position of a shift-pole must represent the average rotation over some length of time. If the time is less than a million years or so the shift-pole is considered to be 'instantaneous', if the time is longer 'stage pole'. The most commonly determined stage poles are those back from the present day to some time in the past, rather than between two moments in the past. For this account the values of most import are the position and rotation rate of the present day instantaneous pole, effectively the stage pole for the last million years. The best way of determining a present rate for New Zealand from plate tectonics is from the sea floor magnetic anomalies from the present day back to about 5 My ago. There will be a stage pole position to go with this rate.

Two different ways have been proposed for finding the position of the instantaneous pole for the Indian-Pacific plate boundary. The first was used by Falconer (1973) who noticed that the epicentres of the earthquakes on the southern part of the Indian-Pacific plate boundary lie on a circular arc. Falconer by assuming that faulting was simply dextral, placed the shift-pole at the centre of the arc. The arc relationship had not been noticed before

because a Mercator projection had previously been used to plot the epicentres, and on Mercators projections circular arcs do not plot as circular arcs in high latitudes. The second way was used by Chase (1978) by a study of world-wide earthquake slip vectors, transform fault trends, and sea floor magnetic anomalies. The Falconer pole (lat 55.8°S long 174.7°E) lies well outside the 95% confidence limits of the Chase pole (lat $62 \pm 2^{\circ}\text{S}$ long $174.3 \pm 2^{\circ}\text{E}$), and indicates that dextral slip is not the only motion on the circular arc. The Chase pole is adopted here. It lies to the south-west of the shift-pole positions inferred from sea floor magnetic anomalies.

The anomalies used to determine the shift-pole positions are those recorded in new ocean crust that represent the spreading between the Indian and Antarctic plates, and between the Pacific and Antarctic plates. The 'composition' or vector sum of the two rates of spreading for any particular time interval will give the position of the stage-pole and amount of rotation for the Indian-Pacific plates for that time interval. The inferred stage-poles for the Indian-Pacific plates show that there has been a southerly movement in the position of the instantaneous pole with time.

The nature of the movement on a plate boundary depends on the direction of the boundary relative to circular arcs centred on the shift-pole, and on the sense of rotation at the shift-pole. In figure 2 the consequences of dextral movement are shown for a plate boundary that does not have a constant orientation relative to circular arcs about the shift-pole. The Indian-Pacific plate boundary through New Zealand also does not have a fixed orientation, and so the sense of displacement is first reverse-dextral for the Alpine Fault, then dextral for the branch faults through Marlborough, then probably normal-dextral through Cook Strait, then reverse-dextral along the east coast of the North Island, and finally reverse along the Tonga-Kermadec Trench (Fig. 2 and 3). According to the rotation rate and distance from the Chase shift-pole the net rates should be about 55 mm y^{-1} in the northern part of New Zealand and about 40 mm y^{-1} in the southern part.

Both active faults and earthquake epicentres can be used to delimit the active belt that is the Indian-Pacific plate boundary in New Zealand. Figure 4 shows that the belt is indicated by the epicentres of historic large earthquakes, with a 'seismic gap' between 42° and 44°S . The consequences of this seismic gap are being actively argued at present, and will be discussed later in this work. The detailed study of small but frequent earthquakes ('micro-earthquake studies') allows the direction of crustal movement at the time of the earthquake to be found for many earthquakes. In the Southern Alps where

large earthquakes have not been observed, Scholz et al (1973) used micro-earthquakes to demonstrate oblique compression on the Alpine Fault, a result that agrees with the nature of movement predicted from the shift-pole and plate boundary orientation.

② The next most direct way of finding the direction and rate of tectonic movement is from the difference between land triangulations and retriangulations. It so happens that horizontal angles can be measured to about 10^{-5} whereas old distance measurements are accurate to about 10^{-4} only. The changes in horizontal angles give information on changes in shape, but not size or orientation. Walcott (1977) has interpreted retriangulation data for much of the country (Fig. 4), and by assuming homogenous strain (that is, uniform deformation in which straight lines remain straight, and parallel lines remain parallel) calculated an average strain rate of $4.5 \times 10^{-7} \text{ rad.y}^{-1}$. The strain rate is equivalent to relative movement of 45 mm y^{-1} across a 100 km wide belt and is the same as that predicted from the Chase (1978) pole.

The rate of tectonic movement can also be found from observed onshore rock deformation. Relative movement of the Indian and Pacific plates involves both horizontal and vertical displacements in the rocks of the intervening belt. The most obvious displacements occur on faults in the belt, and for these, movement rates can be found from the amount of displacement and the time since displacement started. Less obvious displacement may occur as distributed deformation between the faults, as proposed by Walcott (1977), but this cannot be observed directly. In theory the rate and direction of relative movement can be found by vector summing the individual rates and directions across the belt, but in practice only the displacement on faults are available. Non-elastic deformation between faults cannot be determined accurately at present so the measured rate is less than the actual.

Horizontal Rates

It was suggested in 1948 that the structure of the pre-Cretaceous rocks of the South Island would be considerably simplified if a dextral displacement of some 500 km on the Alpine Fault were to be taken out. There is now general agreement among geologists that large dextral displacement has taken place, but no agreement on when. Conservatists spread the displacement over the maximum time interval and make it more or less uniform for the

last 100 My. Radicals confine it to the last 25 My.

A feature that appears to support the conservatists is the apparent 120 km dextral displacement of lamprophyre dikes that are about 100 My old according to K/Ar dating. The features used to infer the 500 km displacement are distinctive rock strata that dip steeply and are likely to extend down for several kilometres. The distinctive rock strata have a well defined strike, and because they extend to depth are ideal reference lines on maps for defining horizontal displacements. The lamprophyre dikes are quite different. They are poorly exposed; little is known about their strike, and almost nothing about the strike of the belt of dikes as a whole. The belt of dikes is the same kind of feature as the distinctive rock strata that have been used as reference lines for the Alpine Fault displacement, and if the strike of the belt could be determined there would be an additional 'fix' on the displacement.

It is suggested here that the strike of the belt of dikes is more oblique to the fault than the strike of the rock strata, and that movement was largely dextral but also partly reverse (narrowing). The reverse component when taken out will be represented by a gap between the two sides of the fault that are now in contact. Neither the size of the gap nor the exact angle between the dike belt and fault are known, but an estimate is shown by an inset to figure 5. It will be noted that the dikes have the effect of reducing the total dextral displacement by about 40 km, and the total displacement is shown as 460 km.

Horizontal displacement on the Alpine Fault has occurred progressively over the last 25 My or so. Reference lines of different ages have been displaced proportionate to their age as is shown by figure 5, and when the displacement of each reference line is plotted against its age (Fig. 9) a well defined line is shown that represents a substantially uniform rate of 20 mm y^{-1} for the last 25 My.

An important factor in the horizontal rate story that is not directly relevant to the main content of this thesis is the bending of the New Zealand region. The bending is dextral (in the same sense as the horizontal displacement on the Alpine Fault) and its extent is shown by the bending of pre-Cretaceous rock boundaries. It is assumed here that bending and dextral faulting started some 60 My ago and that they are still continuing. Bending is by sinistral movement on steeply dipping faults that are at approximately 45° to the dextral faults. The movement during the 1929 Murchison earthquake on the White Creek Fault is assumed to be active faulting related to the

bending. The total displacement from faulting and bending within New Zealand is about 700 km. Offshore to the south-east Christoffel and Davey (pers. comm.) suggest that the Stokes Magnetic Anomaly (a geophysical reference line) can be traced across the Campbell Plateau, but that it has been displaced by the 'Campbell Fault' in a dextral sense. The simplest explanation for the displacement is 3-400 km of dextral movement on a circular fault that passes out through the Bounty Trough and has a centre at lat 56°S long 179°E. With this offshore displacement, total displacement across New Zealand is about 1000 to 1100 km and is the same as the amount predicted from the sea floor magnetic anomalies.

Vertical Rates

✓ Present uplift on the Alpine Fault is partly tectonic and partly glacio-isostatic. The agreement between interglacial and postglacial rates shown below supports Mathews' 1967 conclusion that the total rate is ten times the glacio-isostatic part, and so the glacio-isostatic effects have been neglected.

basal? The Indian-Pacific plate boundary in the South Island has both dextral and narrowing components; the dextral component is expressed as dextral horizontal movement, the narrowing by differential uplift. Early geologists on the West Coast considered that a peneplain had extended directly from the western granite mountains across the fault to the schist mountains of the Southern Alps, but the growing appreciation in the 1950's of the Alpine Fault as a major feature led to an interest in vertical displacements along the fault.

Alpine Heating? As for the dextral horizontal displacements, there is agreement that reverse vertical displacement has taken place, but no agreement on when. A feature that appears to support a conservatist view is a series of K/Ar 'ages' of 6 to 15 My from garnet rank schists immediately to the east of the Alpine Fault. The garnet schists are the most westerly of rocks that show an increase in metamorphic rank from zeolite rank greywackes in the east to the garnet schists at the fault. All the schists are presumed to represent metamorphosed greywackes, with the garnet schists representing the most deeply buried greywackes yet exposed; they recrystallised at depths of 18 km or more.

K/Ar ages are proportional to the ratio of radiogenic ^{40}Ar to the ^{40}K content of the rock. In a closed system all the argon is trapped, and the known decay rate of ^{40}K together with the measured amounts of ^{40}K and ^{40}Ar allow calculation of the time since the system closed to argon loss. Some

Peneplain

other clues
what are the
diagnostic clues for the peneplain?

rocks already contain some ^{40}Ar at the time they become a closed system; these rocks give 'ages' older than the time the system became closed. The schists dated by Sheppard et al (1975) do not show a relationship between ^{40}K and ^{40}Ar ; instead the amount of ^{40}Ar is proportional to the metamorphic rank, and this implies that argon has migrated and that the 'ages' do not date any real event. Sheppard et al (1975) considered that their 'dates' represent late Cenozoic partial degassing of argon trapped since Cretaceous uplift cooled the rock. They suggested that 8 to 12 km of uplift occurred 100-120 My ago, and then a further 3 to 7 km in the late Cenozoic. They considered that the most recent uplift was 'in accord with the evidence that the mid-late Tertiary peneplain in the Southern Alps was only slightly above the present-day highest peaks'. In Otago the peneplain is a few hundred metres above sea level near the coast, but rises to 1800 m inland. Further inland still it cannot be recognised, and is presumed to lie above the present peaks and to have been completely eroded. There is however no evidence other than the summit height accordance of the present-day peaks to suggest that the peneplain was only slightly above the peaks, and as will be discussed below, this accordance is thought by the author to represent a balance between uplift and erosion. The K/Ar 'ages' are most easily explained in terms of recent uplift alone, with most rapid uplift near the fault exposing the highest metamorphic ranks and giving them the youngest closure times. The 'ages' are confounded by partial degassing of schist at depth and the addition of this argon to the argon in the surface rocks.

Radicals suggest that most if not all of the vertical displacement on the Alpine Fault has occurred in the last 2 My or so. They point to the lack of Alpine schist pebbles in lower Tertiary and older conglomerates as evidence against slow uplift, and to the present high uplift rates on the Alpine Fault. Good evidence for rapid uplift comes from displaced reference lines of several types, and evidence for uplift of the eastern Southern Alps comes from deformed lake shorelines there. The evidence is summarized in figure 6 and discussed in detail below.

Although schist has been thrust over moraine at several places along the Alpine Fault, the age and vertical displacement are not well known. Ten other features provide uplift rates.

According to Cooper (Otago University, pers. comm.) a 3.5 ky old faulted river terrace 10 km south of Haast indicates a slip-throw rate of 5 to 6 mm y^{-1} on the Alpine fault with the southeastern side being upthrown. From three to

seven kilometres inland between the Waiatoto and Haast rivers there is a prominent bench that is almost certainly marine-cut at a height of 1500 m (Fig. 7). If the bench is last interglacial (100 ky) then the shift-throw rate is 15 mm y^{-1} ; if penultimate interglacial (230 ky) as thought by Cooper the shift-throw rate is 6.5 mm y^{-1} . Slip-throw is direct fault movement, shift-throw is the direct fault movement plus drag and for the Alpine Fault takes place over a width of three to five kilometres.

At Paringa 50 km to the north-east along the fault marine shells dated by radiocarbon at 13 400 yBP (Suggate, 1968) occur in sediment that fills an old fiord now occupied by the Paringa River. The shells are on the upthrown eastern side of the Alpine Fault and have been lifted about 140 m giving a rate of 10.7 mm y^{-1} .

At Wanganui River 110 km further to the north-east there is a 930 ± 150 yBP old river terrace that is stranded 11 m above the floodplain immediately upstream of the fault (Suggate, 1970). There are no stranded terraces downstream from the fault, and so the rate throw on the fault is inferred to be at least $12 \pm 2 \text{ mm y}^{-1}$. There is also a higher terrace (70 m) immediately upstream of the fault that is 8 ky old if its age is proportional to its height, and so may correspond to the Birch Hill advance on the opposite side of the Southern Alps. The height of the higher terrace relative to the floodplain increases upstream until it is 140 m above 5 km upstream from the fault. Further upstream its relative height decreases.

At Hokitika River a further 35 km north-east along the fault there is a terrace similar to the higher one on the Wanganui but which is exposed on both sides of the fault. On the upstream side it slopes steeply upstream and indicates a slip-throw of 60 m and a shift-throw of at least 120 m (Wellman, 1955). The terrace is probably 10 ± 2 ky and the uplift rate at least $12 \pm 3 \text{ mm y}^{-1}$.

At Haupiri River 75 km further to the north-east and north-east of the point where the Hope Fault joins the Alpine Fault there are stranded terraces on the upstream side and none on the downstream side of the fault. The terraces are undated.

At Maruia River 50 km further to the north-east an undated terrace indicates a throw of 20 m.

At Lake Rotoiti a further 75 km to the north-east stranded lake shorelines indicate a slip-throw of 5 m and a shift-throw of 14 m 3 km upstream from the fault (Wellman, in prep.). Moraine at the lake peninsula has a slip-throw of 9 m. The shoreline is probably 10 ± 2 ky old and the shift-throw rate is $1.4 \pm 0.3 \text{ mm y}^{-1}$.

From the slip-throw and shift-throw rates above I infer that half the uplift on the Alpine Fault is by drag within the Alpine schists immediately to the east of the fault. It is not yet known if the drag is by pervasive rock deformation or by movement on many small, discrete schistosity plane faults.

On the eastern slopes of the Southern Alps the lakes that partially fill glacial troughs have been progressively emptied by downcutting of their outlets, and are now rimmed by stranded shorelines. Some of the better preserved shorelines are tilted and rise higher above the present lake level towards the mountains, suggesting differential uplift of the mountains. Wellman (in prep.) has measured and correlated these shorelines. They indicate a relatively simple pattern of uplift with rates increasing exponentially towards the mountains and the Alpine Fault and so doubling every 7 to 8 km. By integrating the exponential relationship Wellman deduced uplift rates of about 0.5 to 2.0 mm y^{-1} relative to sea level for the lake shores.

To the west of the Alpine Fault uplift rates relative to sea level come from five places: at Lake McKerrow marine silts mentioned by Wellman and Wilson (1964, p.712) that are about 6 ky old give 1.5 mm y^{-1} , but there was no evidence of differential uplift on the fault; at Sardine Terrace 17 km north-east of Haast a 50 m last interglacial marine bench gives 0.5 mm y^{-1} ; at Pug Creek near Franz Josef Glacier young but undated marine shells in silts mapped as last interglacial by Warren (1967) give $0.6+ \text{ mm y}^{-1}$, and 3 km away at Cement Hill a probable last interglacial bench at 220 m gives 2.2 mm y^{-1} ; at Kumara a last interglacial marine bench gives 0.6 mm y^{-1} ; and at Lake Rotoroa stranded lake shorelines (Wellman, in prep.) are 5 m higher near the fault than at the lake outlet and give $0.5+ \text{ mm y}^{-1}$. The rates suggest that the western side of the fault is being dragged up at a tenth of the rate that the eastern side is being dragged up. The uplift on the eastern side extends at least 60 km from the fault to Wellman's tilted lake shores.

The eastern (Pacific, or Southern Alps) side of the fault is being upthrust relative to sea level and to the western (Indian) side. The following formula is designed to fit the pattern of uplift in the central South Island where the pattern is simplest, subject to the constraint that the area under the uplift curve must balance with the 22 mm y^{-1} plate movement of 27 km thick crust:

$$\text{uplift} = 12.89 (x+1)^{0.525} \exp(-0.072(x+1)) \quad (1)$$

where the uplift is in mm y^{-1} for a point x km east of the fault. As seen from figure 10 the agreement between formula and data is good.

An uplift map of the South Island has been produced by Wellman (in prep.) from vertical rate data and from an assumed relation between summit heights, rainfall, and uplift. Uplift rates were inferred to be highest near Mount Cook, but localised centres of rapid uplift also occur in the Kaikoura ranges (Fig. 8).

The rate data is summarized in figure 9. I infer long term horizontal displacement rates of 40 mm y^{-1} , half faulting, half bending, and vertical rates of up to 22 mm y^{-1} , half as slip-throw, the rest distributed. These rates are consistent with the total rate of 45 mm y^{-1} from the sea floor magnetic anomalies for the last 5 My. At the inferred rates the 460 km of horizontal fault movement could be achieved in 23 My and the Southern Alps could be lifted 3 km from sea level in 140 000 y.

The effects of the rapid tectonic displacements are shown in a remarkable sequence of conglomerate bands to the west of the Alpine Fault near Springs Junction (Cutten, in prep.). Pebbles in the bands increase in metamorphic rank upwards from Chlorite I rank greywackes at the base to garnet schist at the top. The sequence is entirely non-marine and is thus poorly dated. Cutten considered the greywackes were derived from Caples Group rocks. The Caples rocks are now 420 km to the south-west of Springs Junction. From stratigraphy the oldest conglomerate band is thought by Cutten to be about 11 My old. To the south near Greymouth the Old Man Gravels about 2 My old are the oldest formation with Alpine schist. It is inferred that 420 km of dextral displacement has taken place in 11 My and that the schist was first unroofed in the Southern Alps about 2 My ago.

Crustal Narrowing, Uplift and Erosion

The cause for the horizontal and vertical displacements within New Zealand is the interaction of the Indian and Pacific plates. As shown in figure 2 the plate interaction results in dextral movement combined with compression on the Alpine Fault and narrowing across the Alpine fault zone.

There are only three ways in which crustal narrowing can occur, by either downward, upward, or sideways movement. Approximately 80 km of narrowing has occurred across central New Zealand in the last 15 My (Walcott, 1977), so assuming a crustal thickness of 27 km and an interaction length of 500 km about $1.1 \times 10^6 \text{ km}^3$ of crust must have been 'lost'.

For oceanic crust the downward, upward, or sideways motion can be taken up by underthrusting, overthrusting or sideways movement, and is then termed subduction, obduction, or transform movement. Extensive underthrusting of continental crust is limited by its buoyancy, although a limited amount of narrowing may be taken up by crustal thickening. It is not possible to accommodate all the narrowing across New Zealand in this way because gravity observations show that the Southern Alps are in isostatic equilibrium, and this indicates only enough crustal thickening to compensate for the mass of the Southern Alps. Taking the Southern Alps to be a prism 3 km high, 80 km wide, and 500 km long, and densities of crust and mantle to be 2.65 and 3.30 t m^{-3} , the total volume of thickening must be five times the volume above sea level, and is $0.3 \times 10^6 \text{ km}^3$ and accounts for 25% of the 'lost' crust.

Continental crust may be 'lost' by being moved sideways and away from the point of collision. Crustal blocks separated by faults can be pushed sideways much as orange pips are squeezed out between finger and thumb. This has been the explanation proposed for continent-continent collisions in Turkey and the Himalayas. In New Zealand Walcott (1977) suggested that crustal blocks separated by faults like those in Marlborough could have moved north-eastwards and added on to the eastern margin of the North Island. The crust lost by narrowing in the South Island would thus be transferred north by additional transcurrent movement and the narrowing taken up by faster subduction of the Pacific plate in the Hikurangi Trench. Progressive southward migration of the southern end of the subduction zone would have allowed the maximum possible narrowing to be taken up in this way. The southeastern limit of the subducted Pacific plate is 600 km from the Hikurangi Trench and at an average rate of 40 mm y^{-1} indicates that the deep part of the plate was near the surface 15 My ago (Walcott, 1977, sec. 4.2). Starting at the same time, the southern end of the subduction zone migrated 460 km to the south-west at about 40 mm y^{-1} before it became trapped against the continental crust of the northern Chatham Rise. The rates and distances indicate that the trapping happened about 3 My ago, and that since then the amount of crust that could be 'lost' in the sideways fashion has been limited.

The third way to 'lose' continental crust involves upward movement of the crust, possibly by overthrusting of one slab of crust over another. Without some other process upthrusting is stopped by gravity. It is certain however that erosion steadily reduces the mass to be upthrust. If erosion proceeds at the same rate as upthrusting it becomes no more difficult to upthrust crust

with time, and upthrusting could continue unchecked. Wellman (1975a) proposed a model for upthrusting by which the lateral narrowing between the Indian and Pacific plates is converted into uplift of the Pacific plate along a curved fault plane (Fig. 10). The narrowing itself is taken up by west dipping reverse faults to the east of the Alpine Fault. Throw on the Alpine Fault decreases to about zero at Lake McKerrow, so there is no narrowing on the fault and the narrowing is assumed to take place offshore, where there are shallow earthquake epicentres, by south-east dipping thrusts that extend around the Fiordland coast. In the north of the South Island some narrowing is assumed to take place immediately to the east of the Alpine Fault, and some much further to the east in the Kaikoura Mountains.

It is the thesis of this work that present day erosion of the upthrust edge of the Pacific plate (the Southern Alps) is about equal to the rate at which the plate is being upthrust. Continental crust 'lost' from the collision zone by erosion is transported away to be deposited on the flanks of the Indian or Pacific plates. The 'lost' crust deposited on the Pacific plate might eventually be uplifted and again eroded. This, the state of affairs for the last 2 My, could continue almost indefinitely. Before the Hikurangi Trench collided with the Chatham Rise crust was probably pushed sideways. A period of about 0.8 My then ensued during which the crust under the Southern Alps was thickened and the Alps lifted to their present height. Since then the upthrusting has been balanced by loss of crust through erosion.

A simple model of crustal narrowing at 22 mm y^{-1} for 27 km thick crust and an interaction length of 500 km gives $3 \times 10^8 \text{ m}^3$ or 800 Mt as the annual volume that must be lost from the collision zone, but because of the thrusts off Fiordland only about 80% or 640 Mt will be expressed as uplift on land. Wellman's uplift map gives an annual volume of $2.3 \times 10^8 \text{ m}^3$ or 610 Mt lifted above sea level in the South Island. The two values agree well, and the adopted value $620 \pm 20 \text{ Mt}$ will be considered later with reference to erosion rates found for the Southern Alps.

The consequences of the tectonic deformation and uplift will now be discussed.

Figure 1. Sketch map of the New Zealand region showing the Indian, Pacific, and Antarctic plates, the three boundaries between them, their triple junction (TJ), and the three shift poles. Also shown are: Tonga-Kermadec Trench (TKT), Hikurangi Trench (HT), Alpine Fault (AF), Southern Alps (SA), and Macquarie Ridge (MR). Conformal projection with 20° grid from New Zealand Atlas.

Figure 2. Diagram showing how the nature of movement at any point on a plate boundary is determined by the azimuth of the boundary at that point relative to its azimuth to the shift-pole; and the nature of movement on the boundary through New Zealand. T: pure transcurrent movement, t: pure tensional movement (spreading), c: pure compressional movement (subduction or obduction), Tt: combined tensional and transcurrent movement ("transtension"), and Tc: combined compressional and transcurrent movement ("transpression").

Figure 3. Map of New Zealand showing major lineations and other important tectonic features. The Alpine Fault (AF) represents the Indian-Pacific plate boundary. To the north-east it splits into the Wairau (W), the Awatere (A), the Clarence (C), and the Hope (H) faults. The Alpine and the other branch faults are either dextral or dextral-reverse. Similar faults occur along the eastern side of the North Island. Also shown is the Taupo Volcanic Zone (TVZ), the important reference lines of probable Permian age defined by the Stokes Magnetic Anomaly (SMA), and the 1 km isobath.

Figure 4. Map of New Zealand showing earthquake epicentres and the relative amount and azimuth of horizontal narrowing (raahn) (units = 10^{-7} rad. y^{-1}) determined from retriangulations as set out by Walcott (1977).

Figure 5. Diagram with sketch figures showing the evidence for the progressive horizontal movement of the Alpine Fault from the difference in displacement of reference lines of different ages. A: Lower Paleozoic ophiolites, B: Stokes Magnetic Anomaly and 'marginal facies' rocks of the New Zealand geosyncline, C: Permian ophiolite belt, D: displaced ophiolite pebble trains, E: displaced river valleys, F: displaced glacial features, G: displaced river terraces, H: geodetic retriangulations, J: geodynamic retriangulations. The offset shown by two areas of Cretaceous lamprophyre dikes (L) when the angle between dike belt and fault is estimated is shown in the inset.

Figure 6. Diagram with sketch sections showing the evidence for the progressive vertical movement of the Alpine Fault from the difference in displacement of reference surfaces of different ages discussed in the text. East side of Alpine fault A: Alpine schist belt, B: Otago peneplain, C: Haast raised beaches, D: Paringa River, E: Wanganui River, F: Hokitika River, G: Haupiri River, H: Maruia River, J: Lake Rotoiti, K: tilted lake shorelines. West side of Alpine Fault L: Lake McKerrow, M: Sardine Terrace, N: Pug Creek, P: Kumara, Q: Lake Rotoroa.

Figure 7. Photographs showing raised marine beaches at c. 1500 m on the upthrown side of the Alpine Fault south of Haast. Top, from north-west; bottom, from south-west.

Figure 8. Uplift map of the South Island from Wellman (in prep.) inferred from tilted lake shores, from a relation between uplift, rainfall and summit heights, and from other data given in the text. The map shows a relatively simple pattern of uplift in the central part of the island, with rates declining exponentially away from the Alpine Fault. Units = mm y^{-1}

Figure 9. Diagrams with logarithmic scales summarizing horizontal and vertical rate data for the Alpine Fault and their associated maximum error bars. Horizontal data from Wellman (in prep.).

The Obduction-Subduction Part of the Australian-Pacific Plate Boundary in New Zealand; the Alpine Fault and the Southern Alps; the Anti-Alpine Fault and the Roots of the Southern Alps

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Figure 10. Scheme proposed by Wellman in 1975 (entire abstract is reproduced) to explain plate interaction in the South Island and the uplift of the Southern Alps. Section is normal to the Alpine Fault. A slab of the Pacific plate 25 km thick is being thrust (obducted) over the Indian plate. The curved fault plane converts the lateral narrowing into uplift. Note anti-fault postulated to subduct the mantle lithosphere. Above the section is the uplift curve according to equation 1 with data from the text identified as in figure 6.

CHAPTER THREE

BEDLOAD ABRASION - SUSPENDEDLOAD ALL IMPORTANT

New Zealand rivers have gravel beds, with gravel that abrades and decreases in size downstream. Suspended load increases with the downstream decrease in size of gravel. For most rivers the suspended load increases from about 50% at the headwaters to about 95% at the coast.

During floods there is a downstream movement of the gravel in a river bed, with fine gravel and sand moving faster than coarse gravel. During the downstream movement the gravel is abraded. In braided rivers the bulk of the river bed is covered only during floods and the gravels on the surface wet and dry many times between floods. Low rank rocks like the Tertiary mudstones of New Zealand fall apart after a few wet-dry cycles, and all rocks are more weathered than they would be if kept permanently either wet or dry. Gravel does not have to move downstream to be abraded. Boulders too large to be moved are abraded by gravel moving over them, and Schumm and Stevens (1973) discovered that where water flow is turbulent the gravel vibrates and abrades in place. For the general case bedload abrasion is probably extremely complex. ✓ For the middle and lower reaches of most of the main South Island rivers the high erosion rate simplifies the problem, and a steady state system results. The amount being eroded from the river headwaters is so large that if gravels of a particular size were to accumulate they would be conspicuous. No such accumulation is known and a sluice-box model is adopted for the middle and lower reaches. Gravel of all sizes goes in at the top end and it all passes out at the bottom end in some form. As it goes through all sizes become smaller, and the smaller sizes travel faster than the larger. Much of the load at the top end is carried in suspension, and as the bedload abrades the proportion and actual amount in suspension progressively increases.

There are two kinds of equipment for simulating abrasion in rivers: tumblers, of the kind used for gemstone polishing; and flumes. Although tumblers do not simulate river action very well they are by far the cheapest to construct and most experiments have been done with them. In a laboratory attempt to determine abrasion coefficients, representative New Zealand rocks were tumbled in a tumbler. River pebbles of various lithologies were weighed, placed in the tumbler drum together, and covered with water. The drum was then rotated for a known number of revolutions and the pebbles reweighed. The distance the pebbles travelled is somewhat less than the movement of the inside of the drum and a 60% value is adopted.

The change in weight or length of pebbles of a particular lithology as they move has been described by the empirical Sternberg law:

$$W = W_0 \exp(-\alpha_W \cdot x) \quad (2a)$$

$$D = D_0 \exp(-\alpha_D \cdot x) \quad (2b)$$

$$\alpha_D = \frac{\Delta \log_e (\text{size})}{\Delta (\text{distance})} \quad (2c)$$

where W , W_0 are final and initial weights; D , D_0 are final and initial diameters; x is the distance travelled in km; α_W and α_D are abrasion coefficients per km for weight and length respectively, and $\alpha_W = 3\alpha_D$. The law applies only to sound pebbles and means that a size-distance plot is a straight line on semi-log paper. The abrasion coefficient (α_D) is proportional to the slope of the line, and is different for different lithologies. As measured in the tumbler the range of α_D for a single lithology is much less than for different lithologies (Fig. 11). As extremes a pebble of Tertiary mudstone looses 50% and a quartz pebble looses only 0.07% of its initial weight after travelling two kilometres.

All pebbles are not sound and the abrasion coefficient decreases appreciably downstream, particularly in river headwaters. This can be shown from rivers with rocks having very different abrasion coefficients. The best New Zealand example is the Clutha where most of the bedrock is quartz veined Chlorite rank schists. My measurements in road cuttings indicate that on average the quartz veins make up 5% of the bedrock. There is a general downstream increase in the proportion of the quartz (a relatively strong rock) to the schist (a relatively weak rock). The ratio by weight of the quartz to the schist and the average size of the five largest pebbles of each lithology were measured at most of the 53 sites. For each site downstream distance is taken as being the channel distance to its furthestmost headwater.

As pebbles become sound with transport, the following law in which the abrasion coefficient decreases downstream according to the reciprocal of the distance is proposed. Figure 12 contrasts the Sternberg law with the one proposed here using data from the Clutha catchment. The data is linearized by plotting $Y = \log_{10}(\text{size})$ against $X = \log_{10}(\text{distance})$ and then fitting the straight line:

$$Y = mX + c$$

$$\begin{aligned} \text{then } \alpha_D &= - \frac{1}{0.4343} \cdot \frac{dy}{dx} \\ &= - \frac{1}{0.4343} \frac{(d(m \log_{10} x))}{(dx)} + \frac{dc}{dx} = \frac{-m}{x} \text{ (point abrasion coefficient)} \quad (3) \end{aligned}$$

The point abrasion coefficient is proportional to the reciprocal of the distance from the headwaters and decreases more and more slowly downstream, finally becoming the Sternberg equation. For a given interval (x_1 to x_2) the interval abrasion coefficient is given by:

$$\bar{\alpha} = \frac{- \int_{x_1}^{x_2} \frac{M}{x} dx}{x_2 - x_1} = \frac{-m \log_e (x_2/x_1)}{x_2 - x_1} \quad (4)$$

which gives the same value found by applying the Sternberg law to the change in size between the endpoints x_1 and x_2 . A coefficient of 'proportional weight change' (α_p) expresses the change in proportion of two lithologies downstream. If each lithology A and B abrades according to the above law, their ratio R changes thus:

$$\begin{aligned} A &= A_0 \exp(-3 \alpha_{DA} \cdot x) & B &= B_0 \exp(-3 \alpha_{DB} \cdot x) \\ A/B &= A_0/B_0 \exp(-3(\alpha_{DA} - \alpha_{DB})x) \\ R &= R_0 \exp(-\alpha_p \cdot x) & \text{where } \alpha_p &= 3(\alpha_{DA} - \alpha_{DB}) \quad (5) \end{aligned}$$

The point and interval abrasion coefficients for the quartz and schist sizes for the 2 km to 294 km interval are calculated from the slope of the lines on figure 12 by equations 3 and 4:

$$\begin{aligned} \text{quartz } \alpha_{Dq} &= \frac{0.171}{x} \text{ km}^{-1} & \bar{\alpha}_{Dq} &= 0.00288 \text{ km}^{-1} \\ \text{schist } \alpha_{Ds} &= \frac{0.574}{x} \text{ km}^{-1} & \bar{\alpha}_{Ds} &= 0.00965 \text{ km}^{-1} \end{aligned}$$

The curved lines corresponding the size changes predicted by applying the Sternberg law have also been plotted to compare with the data in figure 12. The observations indicate that most pebbles in the upstream reaches are unsound and that Sternberg's law is inapplicable.

From equation 5 the interval coefficient of proportional change (α_p) for quartz and schist can be calculated from their interval abrasion coefficients:

$$\bar{\alpha}_p = 3(\bar{\alpha}_{Dq} - \bar{\alpha}_{Ds})$$

$$\bar{\alpha}_p = 3(0.00288 - 0.00965) = -0.0203 \text{ km}^{-1}$$

Figure 13 shows the downstream increase in the weight ratio of quartz to schist gravel. At source the quartz veins are about 5% of the bedrock; at Balclutha 290 km downstream quartz gravel is 55 times as abundant as schist gravel. The source of the schist not being all from the furthestmost headwater explains why many of the data points are below the line. From the slope of the line the point and interval coefficients of proportional change are:

$$\alpha_p = -\frac{1.17}{x} \text{ km}^{-1} \quad \bar{\alpha}_p = 0.0197 \text{ km}^{-1}$$

The interval coefficient so determined is in excellent agreement with the corresponding value calculated above from the change in size of the largest quartz and schist pebbles.

✓ Bedload abrasion by a modification of Sternberg's law thus accounts for all the downstream changes observed in the Clutha catchment.

In their middle and lower reaches the downstream change of largest pebble size follows the simple Sternberg law in many of the rivers studied. The pebbles, having already travelled many kilometres, are sound. For example from the lower Rangitata River the $\log_{10}$ of average size of the five largest pebbles of each lithology plot against distance as straight lines (Fig. 14). The Rangitata, like most other South Island rivers has incised as much as 65 m into an old glacial outwash surface and is also incising at present because the coast is being cut back at 0.7 m y^{-1} (Speight, 1950). Thus the channel has been shortening, steepening, and degrading in the long term and so no upstream accumulation of particular gravel sizes can be causing the downstream change in pebble size.

Abrasion coefficients from the Rangitata and for the other rivers examined are much larger than the coefficients determined experimentally. Figure 15 relates river and tumbler values for many rivers and many lithologies. The scatter is large, but the tumbler values are roughly a tenth the river values. ✓ The difference is attributed to abrasion without downstream movement

for the high rank rocks and to wet-dry weathering for low rank rocks. Only 10% of the total abrasion is caused by downstream pebble movement. Shaw (pers. comm.) has found the same difference for the braided rivers of Alberta and explains it in the same way.

In order to find the sizes produced by the abrasion of gravel, drums two-thirds filled with pebbles of a single lithology were tumbled and the resultant dried and sieved. Marshall (1929) and others have suggested that the main product of gravel abrasion is mud, but I found that dependent on lithology that silt or even sand (from granite) was the main product (Fig. 16). Irrespective of whether the product is mud, silt, or sand, it is sufficiently fine to be carried as suspended load by the turbulent New Zealand rivers.

Thus abrasion transforms bedload into suspended load.

If part of a river channel is flowing entirely in alluvium it is unconfined both vertically and horizontally and should achieve a smooth graded profile between source and sink, the slope at each point depending on the amount of bedload being carried. The shape of the long profile thus provides information on the change in amount of bedload and hence the rate of abrasion of the bedload downstream. The sluice-box model is used for rivers like those of the Canterbury Plains that drain wet steep mountains and then flow across relatively dry plains of river gravel.

Abrasion of the bedload can be described by an abrasion coefficient for weight (α_w) in the same way that the abrasion of a single pebble is described. If x is the distance along the part of the channel considered then:

$$\text{final load} = \text{initial load} \cdot \exp(-\alpha_w \cdot x) \quad (6)$$

If the amount of bedload moved is directly proportional to the channel slope, as will be considered in chapter 4, then (A a constant of proportionality):

$$\text{final load} = A \cdot (\text{final channel slope}) \quad (7a)$$

$$\text{initial load} = A \cdot (\text{initial channel slope}) \quad (7b)$$

The river long profile can be described by a least squares quadratic fit of channel elevation (z) to distance (x) (units = km):

$$z = L + Mx + Nx^2 \quad (8)$$

which can be considered as the Taylor series expansion of an exponential height-distance relationship with the third and higher order terms ignored. The statistical curve fitting should provide a more reliable estimate of

channel slope than is available from simply measuring the distance between successive contours from a map. The channel slope at any point in the segment is the derivative (dz/dx) of the quadratic equation:

$$dz/dx = M + 2Nx \quad (9)$$

with boundary conditions $x = 0$ at start and $x = x$ at finish, and combining equations 6, 7a, 7b and 9:

$$M + 2Nx = M \exp(-\alpha_W \cdot x)$$

or
$$\alpha_W = \frac{-1}{x} \cdot \log_e (1+2Nx/M) \quad (10)$$

Hence a bulk abrasion coefficient can be determined from the long profile, and can be used to determine changes in the bedload transport.

The following example is from the Waiho River of south Westland from the Alpine Fault to the sea.

$$\text{quadratic fit: } z = 0.154 - 0.0114x + 0.000155x^2$$

$$\text{distance: } 17.8 \text{ km}$$

hence
$$\alpha_W: 0.037 \text{ km}^{-1}$$

$$\alpha_D: 0.012 \text{ km}^{-1}$$

Substituting the abrasion coefficient in equation 6:

$$\text{bedload at coast} = 0.52 \text{ bedload at Alpine Fault}$$

so that one half of the total bedload has been lost to the suspended load by abrasion between the fault and the coast. Figure 17 shows these changes and the long profile of the Waiho River. Unfortunately only relative changes in bedload transport can be estimated, but this has been done for 16 segments of rivers in the South Island (Table 1). For rivers with schist bedload the average α_D is 0.0160 km^{-1} , three times the 0.0060 km^{-1} average for greywacke bedload. The coefficients are about one third those found by measuring length changes of the largest pebbles. The differences may be due to large pebbles being abraded more rapidly than small for the distance they move downstream and indicate that as expected the abrasion in place process is more important for large than small pebbles.

The relative importance of bedload and suspended load can be estimated from the abrasion of rock pairs with very different abrasion coefficients:

quartz and schist for the Clutha River; greywacke and argillite for the main rivers to the north. Preferential wear of the softer lithology leaves quartz to dominate the bedload of the lower Clutha and greywacke the rivers to the north. Assuming a source terrain bedrock with an average of 5% quartz, less than 6% of the total load at Balclutha near the mouth of the Clutha moves as bedload (Fig. 18), the rest moves as suspended load. Estimates from Roxburgh dam (Thompson, 1976) indicate that before the dam was built 2.6 Mt y^{-1} was carried by the Clutha as suspended load. The bedload at Balclutha is about 0.1 Mt y^{-1} (Table 2) and so makes up about 4% of the load, in agreement with the analysis from bedload composition.

Experiment and field observations show that bedload is of minor and suspended load of major importance for almost all New Zealand rivers. Hence if erosion rates are to be inferred from river load it is the suspended load rather than the bedload that needs to be well evaluated.

Figure 11. Range of drum tumbler abrasion coefficients for 24 representative New Zealand rocks. Bars represent range of each lithology determined for different pebbles. Total range is from 0.00007 km^{-1} for agate to 0.19 km^{-1} for Tertiary mudstone.

Figure 12. Left: Semi-log diagrams to show systematic difference between fitted straight lines and data points for downstream decrease in size of the five largest quartz, and of the five largest schist pebbles for about 40 streams within the Clutha basin. Right: Log-log diagrams (with normal values) showing same data as at left. Curved lines are those that are straight on the left diagrams, and correspond to a constant abrasion coefficient for the pebbles; new straight lines drawn in by eye represent abrasion coefficients that decrease with distance from source.

Figure 13. Log-log diagram with data points and eye-fitted line showing the "downstream distance from source" increase in proportion of quartz to schist. About 40 streams and rivers represented. At source the quartz is 5% of the bedrock, and at Balclutha 295 km downstream quartz is 55 times as abundant as schist. Each downstream distance is taken as being the channel distance to furthestmost headwater, and the spread of data points below the line is due to much schist not having travelled the full distance.

Figure 14. Semi-log diagram with data points and eye-fitted lines showing downstream decrease in length of pebbles (average of five longest) of four different lithologies in the Rangitata River. The abrasion coefficients given in brackets are determined from the slope of the lines.

Figure 15. Log-log diagram comparing experimental abrasion coefficients with observed river abrasion coefficients. For the rivers the largest pebbles were measured to determine the coefficient. Boxes show range of data values for each lithology. The river values are consistently about ten times larger than the tumbler values.

Figure 16. Diagram showing the grainsize of abrasion products from four different lithologies in the drum tumbler. Initial sample: pebbles. Final sample: smaller pebbles and abrasion products sized above. Abrasion of pebbles produces mainly fine grained products that can be transported as suspended load.

Figure 17. Changes in bedload capacity downstream of the Alpine Fault in the Waiho River of south Westland. Water discharge assumed constant for the reach because catchment area and rainfall are much greater above than below the fault. Changes in the capacity reflect bed slope changes, and the concave-up graded profile and decrease in capacity (given in kt y^{-1} above the profile) indicate that much bedload is being abraded during transport down the channel. A quadratic fit to the long profile by the method described in the text gives an abrasion coefficient of 0.012 km^{-1} .

Figure 18. Relative importance of bedload and suspended load in the Clutha catchment as inferred from changes between bedrock and channel bedload. Quartz, a minor part of the bedrock dominates the bedload, so that most schist, the major part of the bedrock is abraded and transported as suspended load at Balclutha.

Table 1. Table giving downstream change in bedload transport rate and abrasion coefficient from equation 10 for segments of 16 rivers in the South Island. 'Start' and 'finish' points are elevation contours in feet and are given together with the quadratic fit obtained for each segment. The dominant lithology of the bedload is greywacke for all except the Cook, Fox, and Waiho rivers which have schist bedload.

River	Segment length			Coefficients of quadratic fit			α_D
	Start ft	Finish ft	km	L	M	N	km ⁻¹
Waihopai	200	700	18.7	0.150	-0.00809	0.0000170	0.0015
Branch	1100	2100	25.7	0.297	-0.0164	0.000187	0.0110
Waiau	0	700	68.0	0.246	-0.00543	0.0000287	0.0062
Waimakariri	0	900	66.8	0.284	-0.00564	0.0000186	0.0029
Rakaia	0	900	61.4	0.270	-0.00439	-0.000025	≈0
North Ashburton	0	1100	52.0	0.322	-0.00918	0.000062	0.0078
South Ashburton	0	1100	52.8	0.318	-0.00947	0.0000677	0.0089
Rangitata	0	1000	48.2	0.311	-0.00730	0.0000169	0.0017
Orari	0	600	31.3	0.184	-0.00904	0.0000994	0.0124
Hae Hae te Moana	0	700	35.4	0.215	-0.0119	0.000167	0.0470
Opihi (Lwr)	0	300	26.4	0.107	-0.00596	0.0000712	0.0126
Opihi (Upr)	800	1600	18.5	0.181	-0.0125	0.000141	0.0097
Cook	0	400	19.4	0.124	-0.00798	0.0000802	0.0085
Fox	0	500	19.3	0.151	-0.0129	0.000266	0.0274
Waiho	0	500	17.8	0.154	-0.0114	0.000155	0.0124
Buller	1200	1900	16.1	0.213	-0.0184	0.000327	0.0175

CHAPTER FOUR

NORMAL RIVER LOAD

Natural events are considered to be of two kinds: abnormal - large and infrequent; normal - small and frequent. For New Zealand systematic river sampling has been for the last two decades, and with a single exception covers normal events only. Normal events are discussed below. In the next chapter it is shown that the additional effects of abnormal events doubles the inferred normal erosion rates.

Dissolved Load

Dissolved load estimates are based on routine drinking water analyses of river and lake waters made by the Chemistry Division of the Department of Scientific and Industrial Research. Cyclic salts are carried in with the rain from the sea and added to the salts from rock erosion. The salts in sea water are mostly chlorides and the chlorides in the analyses are assumed to be entirely from the sea. Twice the chloride content is thus assumed to represent cyclic salts and is subtracted from the Chemistry Division values. Figure 19 shows the net concentrations and the annual mass transported to the sea for nine regions of New Zealand. Concentrations are much the same in all rivers, and the dissolved load for all parts of the country is thus roughly proportional to the runoff. The total for the South Island is 11 Mt y^{-1} .

Bedload

Most New Zealand rivers have unstable braided gravel channels, and during large floods the gravel moves downstream. The channel may be scoured to depths of 5 to 10 m near bridge piles. In the Kowhai River of Kaikoura (Marlborough Catchment Board, 1969, Appendix 3) scour chains* recorded more than 1 m of scour and 0.6 m boulders were moved during a flood with a return period of 1.2 y. At

*Scour chains are chains attached to concrete blocks and buried in the river bed. The chains are buried vertically above the blocks, but during flood scour the unsupported length of chain falls and lies at the depth of maximum scour. The chain is dug out after the flood to find the scour depth.

Tartare Creek (south Westland) boulders of 20 to 30 t move during floods, and one 7 t boulder identified by the remains of a stairway once bolted to it has moved 0.5 km in 35 years (T. Smith, West Coast Electric Power Board, pers. comm., 1976). Cobbles are thrown out of the water with violence during some floods (Thompson, 1965), and commonly a loud crashing sound can be heard from the banks of braided rivers in floods as the boulders crash into each other. There is no doubt about downstream gravel movement, the problem is: how much.

Formulae for bedload transport are based on theory and on flume studies. The most widely used are the Einstein-Brown, Schoklitsch, and Meyer-Peter-Muller formulae, but they differ from each other and from field observations by a third of an order of magnitude.

Bedload formulae give only bedload capacity - the amount that would be moved if the supply were unlimited and of the correct size - not the actual amount moved. Because their profiles are graded most New Zealand rivers with loose unconfined beds probably transport at bedload capacity. In confined channels - confined either by bedrock or by boulders much too large to be moved - the channel slope is greater than needed to move the available bedload and the capacity is not achieved.

As well as being a function of channel slope, bedload transport is a function of bed roughness, mean flow, and of the style of flow. According to the Einstein-Brown formula (Rouse, 1950, p.797) the transport of well sorted gravel in a channel can be estimated from

$$\phi = 40 \left(\frac{1}{\psi}\right)^3 \text{ for } \psi < 18 \text{ otherwise } \phi = 0 \quad (11)$$

$$\text{and } \phi = \frac{10^6 G_i / cb}{(.67(\sigma/\rho - 1)gd^3)^{1/2}} = 0.115 \times 10^6 G_i / bd^{3/2} \quad (12)$$

$$\psi = (\sigma/\rho - 1)d/yS = 1.65d/yS \quad (13)$$

where ϕ and ψ are non dimensional parameters representing respectively the transport of grains and the grain resistance to transport.

G_i the transport in $g \text{ s}^{-1}$.

σ, ρ the density of stone (2.65) and water (1.00) respectively in $t \text{ m}^{-3}$.

g the gravitational acceleration = 9.81 m s^{-2} .

b, y, d the breadth and depth of the channel and the median grain size respectively in m.

S the slope, fall per unit distance in radians.

When a mixture of sizes is to be moved this equation has often been applied to each grainsize separately, and the transport rates summed to give the total rate. Thompson (1976) however supposed that all the stones were half the size of the largest readily transported.

$$\begin{aligned}
 \text{Then } \psi &= 9; \\
 \text{and } \phi &= 40/9^3 = .0549 \text{ from (11);} \\
 d &= 5.45 \text{ yS} \quad \text{from (13);} \\
 G_i &= (.0549/.115) 10^6 b (5.45 \text{ yS})^{3/2} \text{ from (12)} \\
 &= 6.07 \times 10^6 b y^{3/2} S^{3/2}
 \end{aligned} \tag{14}$$

The assumption is reasonable if the bed is not protected by many over-large pebbles. The total transport rate represents a balance between a low rate for the largest pebbles (moved only in large floods) and a high rate for fine gravel that is moved almost continuously.

By the Mannings formula:

$$Q_i = b y^{5/3} S^{1/2} / n \tag{15}$$

and equation 14:

$$G_i = 6.07 \times 10^6 Q_i S n / y^{1/6} \tag{16}$$

The depth of most braided rivers at all flow conditions is about 0.8 to 2.5 m so that $y^{1/6}$ is about 0.96 to 1.15, say 1.1

$$\text{then } G_i = 5.52 \times 10^6 Q_i S n \tag{17}$$

The total bedload transport will be the integral of G_i for each Q_i times the fraction of time that Q_i is attained. Writing $q = Q_i/Q$ and using $f = \exp(-aq-b)$ to describe the flow duration curve as explained later in this chapter gives:

$$\begin{aligned}
 G_i &= 5.52 \times 10^6 Q S n q \\
 G &= \int_0^\infty 5.52 \times 10^6 Q S n q \left(-\frac{df}{dq}\right) dq \\
 &= 5.52 \times 10^6 Q S n \int_0^\infty q \left(-\frac{df}{dq}\right) dq \quad \text{integration by parts}
 \end{aligned} \tag{18}$$

$$G = 5.52 \times 10^6 Q S n \cdot \frac{1}{a \exp b} g s^{-1} \tag{19}$$

$$Y = 174 \times 10^6 n \cdot S Q \cdot \frac{1}{a \exp b} t y^{-1} \tag{20}$$

Equation 20 relates long term annual bedload capacity (Y) to the channel roughness (n), to slope (S), to mean flow (Q), and to the flow style ($1/a \exp b$). The equation also relates the bedload transport rate to the energy dissipation (SQ) along the channel.

Gravel is carried by rivers only during flows greater than a certain value. This threshold effect must be considered when gravel transport is calculated by integrating flow duration data. Thompson did not take the threshold effect into account and therefore his values are too high. The author has used $f = \exp(-aq - b)$ for the flow duration expression and so includes all the extreme flows but only a small part of the low flows, equivalent to assuming the flow exceeds the threshold value $\frac{100}{a \exp(b)}\%$ of the time.

The rate of gravel extraction for commercial use is normally held to the same as the natural replacement rate, and therefore gives a good estimate of the amount being transported. In most places where estimates are available there is good agreement between extracted and calculated load (Fig. 20). The loads calculated according to equation 20 are therefore considered realistic.

With n taken to be 0.03 for all the calculations, a and b from flow records or $1/a \exp(b)$ estimated from neighbouring rivers, channel slope S measured from topographic maps, Q from flow records, and a field examination of gravel sizes, the value of Y can be calculated. Table 2 gives the calculated bedloads at one or more sites on all the large South Island rivers and the average bedload per square kilometre is mapped for the South Island in figure 21. Rates are highest in the headwaters and lowest near the coast. The highest values are in the Southern Alps.

The calculated bedload for South Island rivers draining directly into the sea amounts to 8.2 Mt y^{-1} , of which half is carried by the short steep rivers of south Westland. An additional 1.1 Mt y^{-1} of bedload is trapped by lakes so that a total of 9.3 Mt y^{-1} is moved. Expressed as rock with a density of 2.65 t m^{-3} this is $3.5 \times 10^6 \text{ m}^3 \text{ y}^{-1}$.

Suspended Load

Suspended load is all important for determining erosion rates (Chapter 3), and provided it is sampled adequately the amount can easily be determined. Almost all suspended load is transported during floods, and an adequate sampling program would take this into account. In fact the actual sampling is inadequate and the remaining part of this section is devoted to an attempt to get rates from the inadequate samples.

New Zealand rivers carry much lower suspended load concentrations than possible. In arid regions concentrations commonly reach 10^5 ppm but in New Zealand rarely exceed 10^4 ppm 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 (Fig. 22). 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 relation 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, generally three times larger, estimates of the suspended load than do regression lines giving equal weight to all points.

To fit the data a slope (e) of 2.3 for the eye-fitted line is adopted for all the South Island except Fiordland. A slope of 1.7 was adopted for Fiordland. The adopted slopes fit the data fairly well for almost all sites. Once the line is fitted the assumed concentration-flow relation can be described by the slope of the line (2.3 or 1.7) and a single point on the line. Because it represents those flood flows that transport most of the load, the concentration at five times the mean flow (C_5) was chosen for the point and for mean flow Q the concentration C_i for any Q_i is

*rating = relation between two variables, here flow and concentration, determined empirically and used to convert ('rate') measurements of one to estimates of the other. The 'loop rating' is like the hysteresis loop in Physics, but is opposite in sense.

$$C_i = C_S (Q_i/5Q)^e \quad (21)$$

Thompson and Adams (in press) give concentration against flow plots for rivers in the eastern Southern Alps. Additional plots are given in figure 23.

On many large rivers and lakes water levels are now recorded automatically every 15 minutes. Lake levels are recorded so that the inflow can be calculated from the change in lake level. The data is stored on magnetic tape and is processed by the TIME DEPENDENT DATA (TIDEDA) computer system of the New Zealand Ministry of Works and Development (MWD). The TIDEDA system converts water level records to flow records and calculates flow duration curves.

A flow duration curve is a graph of the fraction of time f that the flow exceeds z plotted against z . For the present work all the flow duration curves have been normalised to mean flow: $q = \text{flow}/\text{mean flow}$. The theory of continuous variables predicts that in the extreme f decreases exponentially: $f = \exp(-aq - b)$, a and b are parameters. Thus a graph of $\log_e f$ against q tends to a straight line. The actual flow duration curves from New Zealand rivers do follow this form when allowance is made for unreliable water level to flow relations and the short period of records (Fig. 24). The adopted flow duration equation:

$$f = \exp(-aq - b + (a+b-c)/q) \quad (22)$$

uses a third parameter c to fit the flow duration curve at both high and low flows sufficiently well for the suspended load calculations here. The parameters a , b , and c have the following meaning and range:

' a ' is the slope of the distribution at large q on a graph of $\log_e f$ against q , and large values describe records from catchments with more flood storage or seasonal snow storage and consequent slow recession rate from flood peaks. Its range is 0.2 to 0.5.

' b ' is the intercept of the 'straight' part of the distribution curve on the $\log_e f$ axis, and larger values describe records from arid catchments where more storm rainfall is absorbed before storm runoff starts. Its range is 3 to 5.

' c ' is the value of $\log_e f$ corresponding to mean flow i.e. $c = -\log_e f(1)$, and where the flow is exceeded for a third of the time $f(1) = 0.33 = \exp(-1.1)$ and $c = 1.1$. In all the following work c has been taken as 0.7 because the calculations are not at all sensitive to the value of c .

For many rivers in south Westland and some rivers elsewhere there are no flow records, and mean runoff has been estimated and then assumed for this thesis. In south Westland flow records from Hokitika River, Haast River and Ivory Glacier

(lat 43.14°S long 170.92°E) give runoff values of 11.5, 7.8, and 9.8 m y⁻¹ (MWD computer files). Precipitation at the Alpine Fault is about 4 m y⁻¹ and at Ivory Hut about 8 m y⁻¹, less than the runoff, so the glacier is inferred to be shrinking (Anderton, 1976). In accord with the above data runoff of 7.5 m y⁻¹ has been assumed for the western slope of the Southern Alps between the Hokitika and Arawata river.

For calculating the suspended load carried by New Zealand rivers Thompson and Adams (in press) derived the following simple equations from equation 21 and 22 above:

$$G = h C_5 Q \quad (23a)$$

$$\text{and} \quad Y = 31.56 G \quad (23b)$$

$$\text{where} \quad h = \int_0^\infty 5^e (aq^{e+1} + (a+b-c)q^{e-1}) \exp(-aq-b+(a+b-c)/q) dq$$

C_5 = suspended load concentration at 5Q in ppm

Q = mean flow in m³ s⁻¹

G = suspended load transport rate in g s⁻¹

Y = annual suspended load in t y⁻¹

The value of h has been evaluated numerically for some normal values of a , b , and e and graphed in figure 25. The figure also provides a sensitivity test to estimate error bounds on G from uncertainties in the parameters. The value of the method was confirmed by Thompson and Adams (in press, table 3) who showed that the load carried by the rivers that flow into the Roxburgh and Matahina hydro lakes calculated by equation 23b was in excellent agreement with the total sediment trapped.

To estimate suspended load by equation 23b a log-log plot of concentration against flow is required to determine C_5 and e (e.g. Fig. 23), a flow duration curve is required to determine Q , a , and b (e.g. Fig. 24), and figure 25 is needed to find h . The method and equation 23b have been applied to catchments with satisfactory data (Tables 3 and 4). In table 4 'h' is estimated for some other catchments with good concentration but poor flow data.

It has already been shown that the suspended load makes up most of the total load and thus accounts for most of the rock eroded. Average erosion is found by dividing the load by the catchment area contributing load. Within many catchments there is a tenfold or larger change in the rate of erosion, and what is required is not just average values but some way of determining the variation within the catchment.

If there is no rainfall, there can be no runoff, no rivers, and no transport of suspended load. The amount of erosion is thus a function of the runoff. Runoff in turn is related in a fairly straightforward way to rainfall, and runoff equals rainfall less evaporation, groundwater movement being neglected. Evaporation itself depends on temperature and humidity and humidity again on rainfall. Hence for a particular region with temperature and rainfall known the runoff can be reasonably estimated.

The measured runoffs for the catchments on the eastern slopes of the Southern Alps are consistent with runoff doubling for each 12.5 km towards the divide:

$$R_x \propto \exp(-x/18) \propto 2^{-x/12.5} \quad (24)$$

where R_x is the runoff x km from the divide. Coulter in 1973 presented a runoff map for New Zealand having about the same doubling distance. The runoff relationship can be converted to a precipitation relationship by allowing for evaporation and:

$$P_x \propto \exp(-x/22) \quad (25)$$

describes the precipitation P_x in mm y^{-1} x km from the divide.

The shape and position of catchments on the eastern slopes of the Southern Alps relative to an arbitrary runoff contour, here taken to be at 5.7 m y^{-1} , can be described by two parameters L and X . In some places the 5.7 m y^{-1} runoff contour is reached, but in others it is not. Most catchments are approximated by a parallelogram with one pair of sides parallel to the runoff contour and roughly parallel to the divide. If L is the catchment length in kilometres normal to the runoff contour and the catchment runoff R in metres per year is known, then

$$X = 18 \log_e \left(\frac{5.7}{R} \cdot \frac{(1 - \exp(-L/18))}{L/18} \right) \quad (26)$$

is the distance in kilometres from the headwaters of the catchment to the 5.7 m y^{-1} runoff contour.

In the case of the Mararoa (Fig. 26) 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 define the position of the 5.7 m y^{-1} 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. 26). The Southern Alps with its long narrow strip with very high rainfall is thus somewhat unusual in its rainfall regularity.

A universal formula for erosion that seems to apply reasonably well to the Southern Alps has been proposed by Fournier (1960):

$$\log_{10} E = 2.65 \log_{10} (p^2 P) + 0.46 \log_{10} (H^2/A) - 1.56 \quad (27)$$

where E = erosion in $\text{t km}^{-2} \text{ y}^{-1}$.

P = precipitation in mm 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 there is no strong seasonal pattern of precipitation and so p is assumed constant ($p \approx 0.1$). The catchments all have similar topography and so H^2/A is assumed constant. Then

$$E_x \propto P_x^{2.65} \quad (27a)$$

and by eqn 25: $E_x = E_0 \exp\left(\frac{-x}{22} \cdot 2.65\right) = E_0 \exp(-x/8.3) \quad (28)$

where E_0 is the erosion for 5.7 m y^{-1} of runoff. By integration the average erosion E in terms of X and L will be

$$E = \frac{E_0}{L/8.3} \cdot (\exp(-X/8.3) - \exp(-(X+L)/8.3)) \quad (29)$$

or $E_0 = E \exp(X/8.3) \cdot (L/8.3) / (1 - \exp(-L/8.3)) \quad (29a)$

Table 3 contains data for calculating the average erosion E for those catchments on the eastern slopes of the Southern Alps used by Thompson and Adams (in press). The data and equation 29a allow E_0 to be estimated from each catchment. For most large catchments with reliable data the values of E_0 are in good agreement. The average value weighted by catchment area is $18\,000 \text{ t km}^{-2} \text{ y}^{-1}$.

Three small catchments in the Waitaki basin give low estimates of E_0 . 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 ice field and presumably requires special analysis. Two catchments in the Wairau basin give high estimates of E_0 . 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_0 for the Arrow and Shotover relative to the rivers to the north can be explained in terms of lithology - schist versus greywacke. As shown by the tumbler experiments in chapter 3, schist is softer and weaker than greywacke and so can be eroded more easily. The values of E_0 suggest that schist is eroded three times faster than greywacke for the same precipitation.

Erosion can be estimated from the suspended load rates for any point x km from the 5.7 m y^{-1} runoff contour by using the value $E_0 = 18\,000 \text{ t km}^{-2} \text{ y}^{-1}$:

$$E_x = 18\,000 \exp(-x/8.3) \quad (30)$$

✓ Suspended loads for the whole of the South Island are shown on figure 27. There are three kinds of area: the central area where measured loads are high and in agreement with known runoff; the coastal areas where the rates are lower than expected from the runoff; and the western slopes of the Southern Alps where the load is based on estimated runoff. The total suspended load amounts to 265 Mt y^{-1} , most of which reaches the sea.

The normal values discussed above are set out below.

transported as	mass (Mt y^{-1})	% of total
Dissolved load	11	4
Bedload	9	3
Suspended load	265	93
Total	285	100

It has already been briefly mentioned that rivers have an almost unlimited capacity to transport suspended load. The amount transported depends mostly on the amount supplied and not on river capacity. The normal situation with limited supply has been discussed above. The abnormal situation with unlimited supply is discussed in the next chapter.

Figure 19. Map showing annual dissolved load from nine regions of New Zealand. Average dissolved salts from data given for domestic water supplies (Chemistry Division reports 2064, 2076, 2085, 2092 and 2161). Range is small within each district and the values were not weighted for volume of flow. Cyclic salts were excluded by subtracting twice the chloride concentration. For the whole of New Zealand the range (in ppm) and amount is small, and so the annual load from each region in Mt y^{-1} is roughly proportional to runoff.

Figure 20. Log-log diagram comparing gravel extraction rates that result in no channel degradation (from Catchment Board data) with the bedload capacity calculated for the river according to the formula derived in the text. In most places extracted and calculated rates are almost the same, and the formula is thus considered to give reliable estimates.

Figure 21. Map of South Island showing calculated bedload transport rates from table 2 expressed as erosion rates in $\text{t km}^{-2} \text{y}^{-1}$ for the site catchment (that part of the river catchment upstream from the site). In river catchments with more than one site, catchments overlap. The method used to show the rates is explained in the inset. For each major river mouth bedload supplied to the coast is shown in Kt y^{-1} .

Figure 22. Top: Changes in flow, suspended load concentration, and fraction of the load finer than 0.02 mm during the 3-5 May 1976 flood on the Mararoa River. Bottom: Log-log diagram of suspended load concentration against flow showing 'loop rating' during the flood. Same data as above, but solid circles are some additional measurements from other floods. Both figures are from Christian and Thompson (1978)

Figure 23. Log-log diagram showing data and adopted relations between flow and concentration for six South Island sites. The straight lines were fitted by eye keeping the flushing effect in mind. The nominal concentration at five times the mean flow is labelled e.g. Haast River 1800 ppm at $1000 \text{ m}^3 \text{ s}^{-1}$. Six further log-log plots are shown in figure 26.

Figure 24. Semi-log diagram showing relation between flow scaled to mean flow and the probability that the scaled flow will be exceeded (flow duration curve in terms of mean flow). The dots represent measurements and the curves represent equation 22 for A: Lake Ohau inflow $a = 0.50$, $b = 3$; B: Lake Te Anau 0.38 , 3 ; C: Manuherikia 0.24 , 4 ; and D: Wairau at Tuamarina 0.26 , 3 .

Figure 25. Semi-log diagram of the function h in terms of parameters a , b and e . "h" is defined below equation 23 and represents the ratio of the mean transport to the product of the mean flow and the concentration C_5 .

Figure 26. Left: Map of the 33 catchments on the eastern slopes of the Southern Alps listed in table 3, with an enlargement of the two Mararoa Catchments showing how each has been represented by a parallelogram with two sides parallel to the 5.7 m y^{-1} runoff contour. Right: Six log-log plots of concentration against flow set out as in figure 23. (Figures 24, 25 and 26 also appear in Thompson and Adams; in press).

Figure 27. Map of South Island site catchments showing calculated suspended load transport rates from tables 3 and 4 expressed as erosion rates in $\text{t km}^{-2} \text{ y}^{-1}$. Areas for which rates have been estimated have not been enclosed.

Table 2. Table giving calculated bedload transport rates for one or more sites on almost all the large South Island rivers. Rates calculated by the simplified bedload formula derived in the text. Values for R and $1/a \exp(b)$ are estimated if values of Q and a or b are not given.

River	Site	A km ²	R m y ⁻¹	Q m ³ s ⁻¹	S 10 ⁻⁴	a	b	$\frac{1}{a \exp(b)}$	Y kt y ⁻¹	Erosion t km ⁻² y ⁻¹
Aorere	Rockville	720	2.0		18					
Takaka	Coast	870	2.0		16			0.2	86	120
Motueka	Coast	2071	0.9		18			0.2	92	105
Motueka	Woodstock	1750	1.03	56.9	26			0.2	110	55
Wangapeka	Blue Rock	447	1.64	23.2	48			0.2	155	89
Motueka	Gorge	163	1.14	5.9	94			0.2	116	260
Waimea	Coast	800	0.6		27			0.2	58	354
Wairoa	Gorge	435	1.24	17.1	52			0.2	43	54
Pelorous	Coast	880	1.2		15			0.2	93	213
Wairau	Tuamarina	3430	1.09	118	15	0.26	3	0.2	52	60
Wairau	Rainbow	505	1.56	25	120	0.34	4	0.19	177	51
Branch	Recorder	580	1.20	22	66	0.14	3	0.05	85	167
Waihopai	Craiglochchart	764	0.68	16.5	90	0.23	3	0.36	270	460
Awatere	Coast	1550	0.7		47			0.22	168	220
Ure	Coast	130	0.7		82			0.2	169	109
Clarence	Coast	3150	0.7		74			0.2	25	190
Acheron	at Clarence	997	0.7	22	74	0.36	3	0.2	540	171
Clarence	Jollies	440	0.86	12	69	0.28	4	0.14	118	118
Hapuku	Coast	140	0.7		265			0.07	28	64
Kowhai	Coast	55	0.7		110			0.2	86	610
Kahautara	Coast	180	0.7		110			0.2	14	254
Conway	Coast	505	0.7		60			0.2	46	254
Waiau	Coast	3290	0.98	102	27			0.2	70	140
Waiau	Marble Point	2048	1.33	86	47	0.28	4	0.1	143	43
Waiau	Glen Hope	715	1.41	32	73			0.07	147	72
Hope	Glen Wye	696	1.59	35	59			0.1	120	170
Hurunui	Coast	2330/2640	1.0		38			0.1	108	155
Hurunui	Mandamus	762/1072	1.41	48	49	0.44	3	0.1	165	71
Hurunui	L. Sumner inflow	310	1.7	28.3	56			0.11	139	182
Waipara	Coast	700	0.4		55			0.15	124	400
Ashley	Coast	1290	0.12	5	39			0.1	25	36
Ashley	Gorge	520	0.34	5.6	74			0.1	10	8
Waimakariri	Coast	3660	0.95	110	21			0.1	22	42
Waimakariri	Gorge	2460	1.52	120	61	0.38	3	0.13	156	43
Poulter	at Waimakariri		2.0		68			0.13	500	204
Waimakariri	below Hawdon		3.2		78			0.15		337
Selwyn	Whitecliffs	164	0.65	3.4	104	0.17	4	0.15		620
Rakaia	Coast	3200	1.87	190	47			0.11	20	121
Rakaia	Gorge	2640	2.32	194	59	0.50	3	0.10	466	146
Harper	Diversions	323	1.75	18	123	0.30	4	0.10	600	225
Wilberforce	Intake	419	3.19	42.4	86	0.35	3	0.06	71	218
Rakaia	above Wilberforce		3.2		74			0.14	271	646
Ashburton	Coast	1600	0.4		69			0.14		550
Ashburton N	Siphon	340	0.5		129			0.1	73	46
Ashburton S	Mt Somers	610	0.47	9	72	0.4	3	0.1	36	107
Rangitata	Coast	1730	1.64	90	60			0.12	70	114
Rangitata	Gorge	1494	1.94	92	74	0.50	3	0.10	281	163
Clyde	at Havelock		3.2		141			0.10	360	240
Havelock	at Clyde		3.2		141			0.1		750
Orari	Coast	815	0.36	9.3	37			0.1		750
Orari	Silverton	521	0.53	8.8	94	0.3	3	0.15	27	33
Opihi	Coast	2330	0.4		29			0.15	64	123
Opuha	Skipton	485	0.45	9.6	72			0.15	67	29
Opihi	Rockwood	368	0.45	5.3	72			0.15	39	80
Tengawai	Pleasant Point	613	0.45		52			0.15	29	80
Pareora	Coast	540	0.45		52			0.15	36	58
Waihao	McCullochs Br.	440	0.45		30			0.15	31	58
Waitaki	Coast	7000/11000	0.94	326	41			0.15	15	33
Hakataramea	Recorder	899	0.18	5	57			0.04	178	25
Godley	L. Tekapo	770	2.25	55	52	0.37	3	0.1	15	17
Forks	Balmoral	130	0.73	3	108	0.54	3	0.13	200	260
Tasman	L. Pukaki	1120	2.82	100	48	0.45	3	0.09	16	120
Jollie	Mt Cook Stn	139	1.6	7.1	173	0.54	3	0.11	276	246
Tasman	above Hooker		5.7		88			0.09	57	412
Hooker	Ball Hut Br.	122	5.4	21	206	0.33	3	0.15		1244
Twizel	Twizel	261	0.51	4.2	74	0.32	3	0.15	337	2760
								0.15	24	94

Table 2 continued.

Hopkins	L. Ohau	920	2.40	70	25	0.50	3	0.10	91	100
Ahuriri	Benmore	1670	0.59	31	56	0.6	3	0.08	75	45
Ahuriri	South Diadem	566	1.23	22	69	0.32	4	0.06	47	85
Otematata	Pumphouse	670	0.41	8.7	110	0.3	3	0.17	83	123
Kakanui	Kakanui	893	0.2		37			0.1	11	12
Shag	Coast	352	0.2		18			0.1	2	6
Taieri	Outram	4800	0.2		24			0.08	30	6
Clutha	Balclutha	12900/20300	0.84	539	5.7			0.06	96	7
Manuherikia	Ophir	2144	0.22	15	49	0.24	4	0.08	31	14
Lindis	Ford	972	0.16	5	57	0.43	3	0.12	18	18
Hunter	L. Hawea	440	1.8	25	74	0.38	3	0.13	127	290
Makarora	L. Wanaka	720	2.6	60	20	0.32	3	0.16	100	139
Mataketake	L. Wanaka	800	2.4	60	25	0.32	3	0.16	122	152
Arrow	Tobins track	196	0.4	2.5	83	0.46	3	0.11	12	60
Shotover	Bowens Peak	1082	1.1	37	24	0.42	3	0.12	57	52
Dart	L. Wakatipu	1046	1.81	60	33	0.30	4	0.06	63	60
Pomahaka	Burkes Ford	1924	0.39	24	18	0.18	4	0.10	23	12
Mataura	Coast	5360	0.33	56.7	8.5	0.65	3	0.08	20	4
Mataura	Gore	3465	0.27	29.6	15	0.65	3	0.08	19	5
Oreti	Coast	3500	0.3		14			0.08	19	6
Oreti	Dipton	1435	0.3		25			0.08	14	10
Aparima	Coast	1580	0.3		14			0.08	9	6
Waiatu	Coast	3320/8060	2.08	530	23			0.04	254	77
Mararoa	The Key	873/1219	0.72	28	51	0.40	3	0.12	64	73
Mararoa	N. Mavora L.	350	1.10	12	13			0.12	10	28
Upukerora	L. Te Anau		1.4		80	0.38	3	0.13		241
Cleddau	Coast	155	6.1	30	56	0.32	3	0.15	131	848
Arawata	Coast	861	7.5		12			0.2	256	298
Waiatoto	Alpine Fault	431	7.5		10			0.2	106	248
Turnbull	Coast	154	7.5		29			0.2	110	720
Okuku	Alpine Fault	257	7.5		22			0.2	140	545
Haast	Coast	1324	7.5		17			0.25	698	527
Haast	Roaring Billy	842	7.5	201	18	0.20	3	0.25	470	558
Moeraki	Alpine Fault	67	7.5		22			0.2	37	546
Paringa	Alpine Fault	232	7.5		24			0.2	140	600
Mahitahi	Coast	155	7.5		29			0.2	111	720
Makawhio	Coast	119	7.5		21			0.2	62	521
Karangarua	Alpine Fault	363	7.5		32			0.2	288	793
Cook	Coast	267	7.5		51			0.2	337	1265
Cook	Alpine Fault	174	7.5		128			0.2	550	3176
Fox	Alpine Fault	93	7.5		128			0.2	300	3176
Waiho	Coast	~162	7.5		59			0.2	237	1464
Waiho	Alpine Fault	162	7.5		114			0.2	460	2828
Waitangi	Coast	68	7.5		17			0.2	29	421
Whataroa	Coast	~447	7.5		21			0.2	232	521
Whataroa	Alpine Fault	447	7.5		46			0.2	510	1141
Poerua	Coast	~60	7.5		30			0.2	45	744
Poerua	Alpine Fault	60	7.5		67			0.2	100	1662
Wanganui	Coast	~337	7.5		22			0.2	184	545
Wanganui	Alpine Fault	337	7.5		53			0.2	440	1315
Waitaha	Alpine Fault	151	7.5		34			0.2	127	843
Kakapotahi	Alpine Fault	36	7.5		223			0.2	200	5533
Mikonui	Alpine Fault	73	7.5		169			0.2	306	4193
Hokitika	Coast	1041	7.5		12			0.18	212	268
Hokitika	Alpine Fault	571	7.5		34	0.28	3	0.18	430	758
Kokatahi	Alpine Fault	222	6		103			0.2	453	2044
Arahura	Alpine Fault	126	5		69			0.2	144	1141
Taramakau	Coast	~1100	4		20			0.15	220	198
Taramakau	Alpine Fault	994	4		69			0.15	680	685
Crooked	Alpine Fault	440	3.2		74			0.15	260	587
Grey	Dobson	3830	2.75	334	13	0.35	3	0.14	322	84
Ahaura	Gorge	790	3.4	86	35	0.38	3	0.13	205	260
Grey	Waipuna	642	2.4	49	34	0.32	3	0.16	135	210
Buller	Te Kuha	5780/6350	2.0	407	10	0.35	3	0.14	300	51
Buller	Berlins	5400/5960	2.0	376	10	0.37	3	0.13	264	49
Inangahua	Landing	1000	2.3	73	37	0.2	3	0.25	352	352
Inangahua	Blacks Point	234	2.3	17	65	0.14	3	0.35	205	876
Maruia	Falls	980	2.1	65	33	0.4	3	0.12	140	142
Maruia	Alpine Fault		2.3		84			0.15		480
Matakitaki	Mud Lake	857	1.85	50	53	0.2	4	0.09	127	148
Matakitaki	Alpine Fault		2.0		72			0.10		240
Glenroy	Blicks	200	1.5	9.7	104	0.25	4	0.07	39	193
Mangles	Gorge	280	1.0	9	98	0.28	3	0.18	82	292
D'Urville	L. Rotoroa	140	2.0	9	94			0.06	26	190
Sabine	L. Rotoroa	140	2.0	9	94			0.06	26	190
Travers	L. Rotoiti	140	2.0	9	129			0.06	36	260
Mokihinui	Coast	694	2.8	62	26	0.30	3	0.17	140	201
Karamea	Arapito	1640	1.8	93	21	0.34	3	0.15	150	91
Heaphy	Coast	160	1.6		21			0.15	13	83

Table 3. Table giving flow and suspended load concentration data and calculated erosion rates for 33 catchments on the eastern slopes of the Southern Alps (Fig. 26). Symbols defined and values in brackets discussed in text. Table also appears in Thompson and Adams (in press).

Catchments Mapped on Figure 26	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 ₀ 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 Waipapai	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 Wairau-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	-	-	-	-	0.84	1140	33
rest of Mandamus	730	24	1.04	35	16	-	-	-	-	0.84	-	-
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 Harpor	320	17	1.68	20	13	0.30	4	0.35	2300	0.43	1350	17
13 Willerforce	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	0.16	1250	0.02	140	11
18 Forks	130	2.9	0.70	26	26	0.54	3	0.16	(550)	0.02	140	(2)
19 Jollie	139	7.1	1.61	12	17	0.54	3	0.16	1600	0.64	5210	(6)
20 Hooker	122	21	5.43	8	-3	0.33	3	0.60	-	-	-	-
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	-	-	-	-	-	-	-

Table 4. Table (similar to table 3) giving flow and suspended load concentration data and calculated erosion rates for a further 23 sites on South Island rivers. Values in brackets estimated from neighbouring rivers, others taken from MWD data.

River	Site	A km ²	R m y ⁻¹	Q m ³ s ⁻¹	a	b	h	C _S ppm	Y Mt y ⁻¹	E t km ² y ⁻¹
Karamea	Arapito	1640	2.16	112	0.34	3	0.56	270	0.53	325
Mokihinui	Burkes	694	3.55	78	0.30	3	0.80	560	1.1	1589
Buller	Te Kuha	5782/6345	2.05	412	0.35	3	0.52	800	5.4	930
Buller	Woolfs	3988/4550	1.62	234	0.37	3	0.45	1150	3.8	958
Buller	Longford	851/144	1.57	70	0.4	3	0.36	650	0.52	607
Inangahua	Landing	1000	2.30	73	0.2	3	2.7	360	2.2	2240
Mangles	Gorge	280	1.00	8.8	0.28	3	1.0	85	0.02	84
Matakitaki	Mud Lake	857	1.84	50	0.2	4	1.1	2100	3.65	4253
Glenroy	Blicks	200	1.56	9.9	0.25	4	0.60	(3000)	0.56	2812
Maruia	Falls	980	2.09	65	0.4	3	0.36	1900	1.40	1431
Inangahua	Blacks Pt	234	2.02	15	0.14	3	8	70	0.26	1132
Grey	Dobson	3830	2.59	314	0.35	3	0.52	350	1.8	470
Grey	Waipuna	642	2.44	49.6	0.32	3	0.66	140	0.14	225
Ahaura	Gorge	790	3.44	86	0.38	3	0.41	650	0.72	915
Orari	Silverton	521	0.53	8.8	0.3	3	0.8	1150	0.26	490
Otematata	Pumphouse	668	0.27	5.8	0.3	3	0.8	130	0.02	29
Pomahaka	Burkes	1920	0.30	20.5	0.18	4	1.5	160	0.16	80
Lindis	Ford	972	0.16	5	0.43	3	0.30	240	0.01	12
Cledau	Milford	155	6.1	30	0.32	3	0.45*	100	0.04	275
Haast	Roaring Billy	842	7.5	201	0.20	3	2.7	1800	31.0	37000
Hokitika	Colliers	332	(7.5)	128	0.28	3	1.0	2800	7.0	21000
Taramakau	Jacksons	994	(4.0)				(1.0)	1500	6.0	6000
Manuherikia	Ophir	2141	0.22	15	0.24	4	0.52	400	0.09	45

*e = 1.7

CHAPTER FIVE

ABNORMAL RIVER LOAD

It has already been mentioned that river transport is limited mainly by the supply in their headwaters. The normal situation is low supply and low transport rates and the abnormal situation high supply and high transport rates. The normal situation was discussed in the last chapter. The abnormal situation follows earthquakes and large floods and its importance relative to the normal is discussed below.

The supply, both normal and abnormal, is mostly from the collapse of unstable hillslopes into the channel. The New Zealand flora has few tap rooted trees to tie the soil to the rock below so the vegetation does not prevent hill-slope instability caused by rock weathering and stream undercutting. After a hillslope has collapsed it continues to erode until the vegetation re-establishes itself and the slope becomes stable again. The cycle of weathering, collapse and revegetation is then repeated.

Collapse of a hillslope may be spontaneous such as the 1908 landslide in the Buller River described by Henderson (1937, p.93). More commonly collapse is triggered by some large, infrequent event such as an earthquake or a large storm. Whether a given hillslope collapses during an event or not depends on the stability of the slope and the magnitude of the event (Fig. 28), and so the effects of any event depend partly on its size and partly on its timing relative to previous events.

The hillslopes collapse as landslips (hereafter called 'slips'), most of which are only a few metres thick and which consist of soil, weathered rock and a few cracked and weathered boulders. From my own observations I estimate that 95% of slipped debris is transformed into suspended load immediately downstream of the slip and 5% remains in the channel to be moved as bedload.

The two main causes of slips are high intensity rain storms and earthquakes. Storms are classified according to their intensity, and in general the greater the intensity the more the slips. 'High intensity' is used here for storms that occur more rarely than a specific frequency, and not for storms in which rainfall exceeds a specific value. Thus high intensity storms as defined occur everywhere and not merely in the areas of highest rainfall. Very few storms have been studied numerically in terms of rainfall intensity and erosion. Two spectacular storms are described below.

The storm of 26-27 Dec, 1957 in the Southern Alps, being widespread was unusually well documented. It covered the northern two-thirds of the South

Island and followed two months of frequent and widespread rain (Hydrology Annual, 1957, p.100-105). In Westland all rivers ran high during November and December, but during the flood all rivers from the Karamea to the Whataroa reached near record levels. Little damage was done except on the Otira River where the river bed aggraded by 7 m and destroyed the railway line for a kilometre.

The storm also caused rivers from the Hurunui to the Waitaki draining the eastern slopes of the Southern Alps to flood. Peak rainfall intensities of 490 mm in 24 hours were recorded at the Hermitage, Mt Cook, and 230 mm in 5 hours in the Rangitata catchment.

The effects in the Godley Valley near the most intense rainfall were described by Scott (1963). Although the storm caused some new slips, the main effect was to wash accumulated fine gravel from the scree slopes into the stream channels. Scott described channel deposits 4.5 m thick and containing 7 to $8 \times 10^5 \text{ m}^3$ of gravel from catchments of 15 km^2 . When he visited them the deposits were being rapidly eroded and transported into the main valley. The accompanying suspended load was doubtless substantial.

From common ages of 30 to 50 years and rare ages of 100 to 120 years for vegetation on older deposits similar to those formed in 1957 Scott concluded that storms like that in 1957 occur every 40 years or so and that some vegetation survives the aggradation. There had been three older storms that had each deposited about 1.5 m of gravel and been followed by a period of soil formation.

The storm of 26 Dec, 1962 in the Rimutaka Range, with peak rainfall intensities of 200 mm per 24 hours was localised, causing slips over a forested area of 40 km^2 drained by the Orongorongo River (Cunningham and Arnot, 1964). The rainfall intensities might be expected once every ten years. Streams draining the area were still muddy three weeks after the storm. Vegetation in many stream channels had been buried under 1 to 3 m of fine gravel, and aggradation extended down the Orongorongo River halfway to the sea. It is noteworthy that slipping and erosion took place under forest cover that had been considered by the Forest Service as satisfactory for preventing erosion.

For earthquakes, massive surface slipping at the Arthur's Pass, Murchison, and Inangahua earthquakes was described by Speight (1933), Henderson (1937), and Lensen and Suggate (1968) respectively. They gave graphic descriptions and photographs but no estimates of the total amounts slipped. Quantitative estimates for the three earthquakes are attempted below.

Slips that show clearly on airphotographs were plotted onto 1:63 360 maps which have a 1000 yard grid. The work was made easier because the maps

had been constructed from the photographs. In New Guinea it was found by Simonett (1967) that there is a rough relation between the map area of a slip and its volume. The conditions in the part of the South Island where the three earthquakes occurred are similar to those in the part of New Guinea where the earthquake occurred, and as far as the writer can judge from rough numerical estimates the relation applies in New Zealand. The relation from Simonett's figure 4.2 is:

$$\log_{10}(\text{volume in m}^3) = 1.41 \log_{10}(\text{map area in m}^2) - 1.07 \quad (31)$$

Altogether there are some 6100 slips for the three earthquakes measured and the following method was used to evaluate the volume for each earthquake.

The relationship above was used to construct a series of templates each of which corresponded in area to volumes on a regular logarithmic scale. The templates were used to size all slips in 5000x5000 yard squares (21 km^2) from the slipped area, and the total volume per square then calculated. The results are summarized in figure 29 and table 5 and outlined below.

The Arthur's Pass earthquake of March 1929 had a magnitude of 6.9. Speight (1933) four years after the earthquake found extensive slips in an east-north-east elongate ellipse from Arthur's Pass to the South Hurunui River mostly drained by the Waimakariri River. Speight's photographs show slips covering 30 to 50% of the hillslopes. The largest fell 600 m, contained 10 tonne boulders, and formed a small lake. Other slips partially blocked rivers and streams. The airphotograph distribution of slips agrees with that described by Speight. A total of $59 \times 10^6 \text{ m}^3$ (106 Mt at a density of 1.8 t m^{-3}) was moved from 820 km^2 , an average loss of $130\,000 \text{ t km}^{-2}$. The most slipped 21 km^2 square lost eight times as much as the average. No downstream effects were reported but gravel from the slips doubtless aggravated a long standing aggradation problem in the Waimakariri River near Christchurch.

The Murchison earthquake of June 1929 in West Nelson with a magnitude of $7^{3/4}$ is the largest that has occurred in the South Island this century. Surface faulting occurred at White Creek (2.2 m sinistral slip, 4.5 m uplift at lat 41.78°S long 172.18°E), but the centre of the slipped area was 40 km to the east-north-east at lat 41.45°S long 172.4°E . The slipped area covers 6000 km^2 and is an ellipse about twice as long as it is wide and which trends 010° (Fig. 29). The total slipped is $1.3 \times 10^9 \text{ m}^3$ (2.3 Gt), an average loss of $380\,000 \text{ t km}^{-2}$. The most slipped 21 km^2 square lost eleven times as much. Major slips dammed the Buller, Maruia, Matakitaki, Matiri, Mokihinui and

Karamea rivers and many of their tributaries. The largest rivers cut through the slips that dammed them in a few days, but smaller rivers such as the Stanley (15 km north of Cobb) are still dammed and about seven substantial new lakes were formed. The enormous amount of weathered rock supplied to the rivers caused extensive river bed aggradation, amounting to 4 m in the Karamea.

The Inangahua earthquake of May 1968 had a magnitude of 7.1 and occurred to the west of the area that slipped in the Murchison earthquake. The total slipped is $52 \times 10^6 \text{ m}^3$ (94 Mt), an average loss of $70\,000 \text{ t km}^{-2}$ from the slipped area of 1300 km^2 . The most slipped 21 km^2 square lost nine times as much as the average. The largest slip had a volume of $4.3 \times 10^6 \text{ m}^3$ and temporarily dammed the Buller River. Three-quarters of the slip was washed away in the first few months. An estimate of the rate at which the slipped volume was moved to the sea is made later in the chapter.

An area of older, almost completely revegetated slips was observed while plotting the slips caused by the Murchison and Inangahua earthquakes. The area is an elongate ellipse between the Little Wanganui and Buller rivers that is partly over the Tasman Sea and which is centred on Granity. As shown by figure 29 the ellipses for the Granity, Murchison and Inangahua earthquakes overlap only slightly. It will be seen from table 5 that the magnitude of the Granity earthquake was 7.4. The earthquake is estimated from the degree of revegetation of the slips to have taken place in the early 1800's. There is no historic record and it probably occurred before 1860.

The many pre-1929 slip-dammed lakes in the Buller-Karamea area (Henderson, 1937, p.93-4) indicate earlier earthquakes. The largest lake, Matiri lat 41.65°S long 172.33°E is dammed by a slip of $1.5 \times 10^7 \text{ m}^3$ that fell from the east side of the valley and mounded up on the west side. The lake was 30 m deep but is now only 20 m deep because it has been partly filled by sediment. Over-topping of the slip by the newly formed lake caused a large mudflow to move down the Matiri Valley. The mudflow moved 8 m sandstone blocks for 2.5 km and these are now mixed with wood from trees that were overwhelmed. A sample of small branches from the mudflow gives a radiocarbon age (sample M29/f20, NZ4230A) of $290 \pm 60 \text{ yBP}$.

Yet older earthquakes are inferred from displacements preserved in faulted river terraces at many places along the Alpine and other major faults in the South Island. At Branch River Lensen (1968) by assuming the highest aggradation terrace to be 18 ky old inferred earthquakes every 900 years on average. A more likely age for the surface is 12 ky however, so that the frequency may well

be once every 600 y. The average displacement of 6 m suggests the earthquakes were of magnitude $7^{3/4}$ at least.

If major earthquakes take place every 600 years and if they cover different areas, then there will be time for considerable weathering between earthquakes and each major earthquake will cause major slipping.

Mud makes up most of the slips and would have been rapidly transported out to sea. Aggradation was reported in the Karamea river near the coast a few months after the Murchison earthquake. In the following decade most of the aggradation was washed out to sea. The suspended load would doubtless have been much greater than the bedload and would have been carried more rapidly.

For the Inangahua earthquake suspended load concentrations taken downstream of the slipped area before and after the earthquake are interpreted as indicating that half the abnormal load supplied by the earthquake entered the sea in the first 19 months. The concentrations were measured from the Buller River at Te Kuha at various times since 1965. Twenty-four measurements have been made, eight before the earthquake, but seven were made at flows less than half the mean and are useless for calculating the suspended loads. To compare the useful concentrations each was converted to a nominal concentration at the standard flow of five times the mean (C_5 at $5Q$, as in chapter 4) by equation 21 with $5Q = 2040 \text{ m}^3 \text{ s}^{-1}$ and $e = 2.3$. Figure 30 shows the nominal concentrations plotted against time with arrows to indicate if the value must be increased or decreased to allow for the flushing effect. The fitted relationship corresponds to a normal C_5 of 1000 ppm, a sudden rise to 10 000 ppm after the earthquake, and an exponential decay to the normal value after eight years. The change in concentration with time after the earthquake is:

$$C_5 = 10\,000 \exp(-0.288 t)$$

where t is the elapsed time in years. The mean concentration for the eight year period is found by integration:

$$\begin{aligned} \bar{C}_5 &= \frac{1}{8} \int_0^8 10\,000 \exp(-0.288 t) dt \\ &= \frac{10\,000}{-0.288 \times 8} (\exp(-0.288 \times 8) - \exp(0)) \\ &= 3900 \text{ ppm} \end{aligned}$$

By equation 23b in the last chapter the pre-earthquake average suspended load at Te Kuha with $h = 0.52$ was:

$$31.56 \times 10^6 \times 412 \times 0.52 \times 1000 = 6.7 \times 10^{12} \text{ g y}^{-1}$$

$$\text{for eight years} = 54 \text{ Mt}$$

For the eight years after the earthquake the load was:

$$8.0 \times 31.56 \times 10^6 \times 412 \times 0.52 \times 3900 = 211 \times 10^{12} \text{ g} = 211 \text{ Mt}$$

The abnormal load carried was therefore about 160 Mt, compared to the 94 Mt that slipped during the earthquake. About half the abnormal load passed Te Kuha and entered the sea in the first 19 months when the concentrations were highest. The result agrees in general with that of Pain and Bowler (1973) who concluded that half the $27 \times 10^6 \text{ m}^3$ that slipped during the Madang earthquake in New Guinea reached the sea in the first six months. The abnormal load carried after the Inangahua earthquake was half as large again as the amount slipped, and the difference probably due to erosion of the bare slip faces.

Abnormal events are catastrophic and cause an enormous amount of erosion, but occur infrequently. Normal events occur frequently but cause little erosion. Erosion has already been estimated for normal events and an attempt is made below to estimate it for abnormal events. The estimate is made by finding the recurrence interval of abnormal events of all sizes (Fig. 31), the erosion caused by each, and then summing their products.

For large storms recurrence intervals are known, but the other data is lacking and it is impossible at present to estimate their contribution to total erosion.

For large earthquakes an estimate of recurrence interval can be made by considering the accumulation and release of strain* as surface movement (Sm) across the whole of New Zealand. From plate movement at 50 mm y^{-1} along the 1000 km plate boundary the surface movement amounts to $50\,000 \text{ m}^2 \text{ y}^{-1}$. Surface movement is released by the faulting during earthquakes, the amount (m^2) being proportional to the product of the fault length and average displacement. For the Murchison earthquake the average displacement was 5 m and the length 100 km, so the release was $500\,000 \text{ m}^2$. Values for the Inangahua and 1855 earthquakes are given in table 5. For earthquakes of magnitude (M) less than 6.4 Gibowicz

* Strain = 'Any definite change of volume or figure exhibited by a solid or liquid mass' in the Shorter Oxford Dictionary physics sense.

and Hatherton's 1975 estimated values for seismic moment calculation were used, and the following composite relation obtained for reverse wrench faults in New Zealand:

$$\log S_m = 1.27M - 4.3 \quad (32)$$

In 1855 a magnitude 8 earthquake occurred near Wellington, but since then there has been no earthquake larger than $7^{3/4}$ in New Zealand. From a relation like equation 32 Wellman (1975b) showed that only half of the surface movement accumulated since the 1855 earthquake has so far been released by earthquakes. For the 1855 earthquake a magnitude of 8.0 was estimated by Eiby (1971), presumably from isoseismals. The 1855 earthquake lifted Cape Turakirae by 2.5 m. Four previous earthquakes in the last 6500 years lifted Turakirae by 3 m, 6 m, 9 m, and 6.5 m (Wellman, 1967), and from their estimated surface displacements are thought to have ranged in size from magnitude 8.0 to 8.4. Turakirae establishes that earthquakes larger than magnitude 8 have occurred in New Zealand. The surface movement doubles for each quarter magnitude step, and hence the few large earthquakes are all important in determining the release of surface movement.

Eiby (1971) gives frequency-magnitude data for earthquakes shallower than 100 km between 37°S and 47°S which includes the 'seismic gap' between 42°S and 44°S , frequency (F) being per year:

$$\log F = -0.89M + 5.59 \quad \text{for half magnitude classes} \quad (33a)$$

$$\log F = -0.89M + 4.89 \quad \text{for tenth magnitude classes} \quad (33b)$$

The product of F and S_m gives the annual surface movement released by earthquakes of a given size. Summing by half magnitude classes to $M = 7.0$ and thence by tenth magnitude classes, half the $50\,000 \text{ m}^2 \text{ y}^{-1}$ surface movement is released by earthquakes smaller than $7^{3/4}$ and half by earthquakes between $7^{3/4}$ and 8.5. The result is consistent with the largest earthquake at Turakirae being about $M = 8.4$ and with rough estimates of the frequency of the Turakirae-sized events in the New Zealand region, and with the accumulation and release of surface movement in the last 120 years. Wellman (1975b) considered that a future large earthquake would occur in the 'seismic gap' to release the accumulated surface movement.

A further relation between F and the slipping caused (T) is taken from data in table 5:

$$\log T = 2.14M - 7.2 \quad (34)$$

There is no reason to suppose that the same relation applies to earthquakes on both normal and reverse faults, and it is considered below only with respect to the reverse wrench faults of the South Island. The annual erosion (E_a) caused by events of size M is:

$$\log E_a = \log F + \log T = 1.25M - 1.61 \quad (35)$$

which can be summed by half magnitude classes to $M = 7.0$ and thence by tenth magnitude classes to 8.5 to give a total erosion rate of 760 Mt y^{-1} . Taking half the 1000 km boundary to be in the South Island, the amount there is 380 Mt y^{-1} . Half the erosion is by the largest few earthquakes, and as neither their frequency nor their effects are well known at the moment, the total could be in considerable error. The adopted abnormal erosion due to earthquakes is $400 \pm 200 \text{ Mt y}^{-1}$.

To the above rate from abnormal earthquake events must be added the rate of 285 Mt y^{-1} from normal events discussed in the last chapter, for a total of about $700 \pm 200 \text{ Mt y}^{-1}$. The total is twice the normal rate.

Figure 28. Diagram showing how slopes tend to become progressively less stable with time and ultimately collapse, thereby becoming suddenly more stable. Collapse that would have occurred anyway is generally triggered prematurely by some abnormal event such as a large storm or large earthquake. The effect of the event depends partly on its size and partly on the length of time since the last event triggered collapse. For example 'A' a slope may not collapse in a major earthquake if a preceding minor earthquake has just caused total collapse of the slope and restored its stability. (Modified after Schumm, 1977, figure 4-16).

Figure 29. Map of central part of New Zealand showing the elliptical area of slipped country caused by five earthquakes discussed in the text. Graphs show on semi-log scales the relation between ellipse area and magnitude and between mass slipped and magnitude.

Figure 30. Semi-log diagram showing the change in nominal concentration at five times mean flow (C_5) at Te Kuha on the Buller River before and after the Inangahua earthquake, and the relation fitted to them.

Arrows show the sense of correction required to allow for these changes.

Figure 31. Semi-log diagram showing recurrence intervals for (Top) large storms at Otira (lat 42.83 S long 171.57 E) and for (Bottom) earthquakes of magnitude eight or less by half magnitude classes for shallow earthquakes in New Zealand (after Eiby, 1971).

Table 5. Table summarizing data from the five earthquakes in the central part of New Zealand discussed in the text and shown in figure 29.

	Granity	1855	Arthur's Pass	Murchison	Inangahua
Date yyyymmdd	?1820	185501	19290309	19290617	19680523
Magnitude	≈7.4	8	6.9	7 ³ / ₄	7.1
displacement maximum m		12		?10	2.4
average m		6		5	1
displaced length from slips km	50	120	30	100	30
surface movement m ²		750 000		500 000	30 000
<u>ellipse of intense landslips</u>					
azimuth	030	035	065	010	060
centre lat °S	41.6	41.0	42.8	41.45	41.83
long °E	171.8	175.4	171.85	172.4	171.97
long axis km	75		35	140	45
short axis km	30		15	45	25
area km ²	1800	≈6000	400	5000	900
<u>amount slipped</u>					
average over area kt km ⁻²			130	380	70
maximum kt km ⁻²			1040	4180	630
total Mt			106	2300	94

On its way to the sea sediment can be trapped in depressions of various kinds. The sediment in the South Island traps provides a rough check on the estimates of river load given above. In order of decreasing age the traps discussed are: growing tectonic depressions; glacial features about 12 ky old - fiords, lakes, and over-widened valleys; postglacial fans; dissected aggradational surfaces; and man-made lakes: they are located on figure 32. Traps differ in their effectiveness; deep lakes are fully effective, but tectonic depressions such as the Canterbury Plains trap only the coarsest part of the load.

Of the growing tectonic depressions the only one providing useful information is the Canterbury Plains. Gravels are 700 m thick near Ashburton (Atkins and Hicks, 1977). Deposition is assumed to have begun with the uplift of the Southern Alps some 2.5 ± 0.5 My ago, giving a subsidence rate of $0.3 \pm 0.05 \text{ mm y}^{-1}$. A Holocene subsidence rate of $0.3 \pm 0.1 \text{ mm y}^{-1}$ has been determined by comparing Suggate's 1968 age-depth values for shells and wood from Christchurch with Holocene values from Victoria (Gill, 1968) and Brazil (Delibrias and Laborel, 1971), Victoria and Brazil being assumed stable. In calculating a volume, the gravel thickness is assumed to decrease outwards in all directions from a 700 m maximum depth. About 10^{12} m^3 of gravel have accumulated over 2.5 ± 0.5 My or $0.8 \pm 0.2 \text{ Mt y}^{-1}$ using a gravel density of 2.1 t m^{-3} . The average accumulation rate for the plains is 0.1 mm y^{-1} , and since the gravels represent only a small part of the total eroded, probably less than 5%, an erosion rate of at least 1 mm y^{-1} is indicated for the adjacent part of the Southern Alps.

During the last glaciation glaciers in the Southern Alps re-occupied valleys that they had cut during the previous glaciation, and are presumed at that time to have removed all sediment from the valleys. The limit of glaciation reached before the glaciers again retreated at about 12 ± 2 ky is marked by terminal moraines and is well known (Suggate, 1965) for most of the South Island; in south Westland however the limit of glaciation was offshore from the Wanganui River south, and the terminal moraines are 100 to 200 m below present-day sea level (Norris, 1977). In many places the glaciers retreated more rapidly than sediment could fill the trough left behind the terminal moraine and fiords and lakes were formed. Elsewhere the valleys were over-widened.

With the exception of lakes McKerrow and Moeraki all fiords north of Milford Sound have been overfilled and now spill sediments. In longitudinal profile the fill is composed of a basal rectangle of depth z superposed by a triangle which thickens from zero at the coast to the known floodplain elevation at the Alpine Fault and has an average height y . At the mouth of the Waiho River depth z is 240 m and at the Poerua River midway between fault and sea 110 m, both values being from petroleum exploration wells (N.Z. Geological Survey open files PR529 and PR528). At the Haast depth z is 125 m at the bridge (Garrick, unpublished Geophysics Division report, 1957) and 200 m at the fault (Garrick and Hatherton, 1974), at Whataroa River 230 m (op cit.), all three values being from seismic refraction surveys. At Paringa a depth z of 100 m is estimated from marine shells (Suggate, 1968). In general the bigger rivers have the greater depths and in table 6 which gives data for establishing the volumes and masses accumulated, the depths are adopted according to the size of the rivers: 230 m for the large and 100 m for the small. The accumulations in each fiord have been expressed as erosion rates for the catchment upstream of the Alpine Fault. The rates decrease from $5 \pm 1 \text{ mm y}^{-1}$ near Mount Cook to about a tenth to the north and south. The rates are minimum values because the traps are now overfilled and because all suspended load would have been trapped only during the earliest stages of infilling. A total of 420 Gt is trapped, which includes eight river channels north of the Wanganui that were treated as filled fiords.

There was a major earthquake in Fiordland in 1826, and in Milford Sound a core from the 8 km^2 Sterling Basin has a thick sand layer overlain by 1.26 m of silty sand. The silty sand is being deposited now but the thick sand layer was probably laid down as a slump deposit during the 1826 earthquake (Pantin, 1964). The silty sand is considered to be suspended load from the rivers, to have a density of 1.25 t m^{-3} and to represent an accumulation rate of 0.1 Mt y^{-1} . The rate gives an estimated erosion rate for the 512 km^2 catchment of $200 \text{ t km}^{-2} \text{ y}^{-1}$ or 0.07 mm y^{-1} of rock, in fair agreement with the estimate (Table 4) of $275 \text{ t km}^{-2} \text{ y}^{-1}$ for the Cleddau River in Chapter 4.

The glacial lakes within glacial troughs are efficient traps and could provide accurate estimates on erosion rates. What is required is dated time planes through the length of the sediment accumulation in the lake to give filling rate. There are two parts to the fill: the delta and the bottom fill. The delta contains both bedload and suspended load, the bottom fill contains mainly suspended load. Little is known about the thickness of the delta deposit and the rate of delta advance, and unfortunately almost nothing definite about

the rate of bottom filling. In the upper Tasman Valley a seismic refraction survey was made in 1971 to find the thickness of the Tasman Glacier. It showed 400 to 500 m of gravel resting on bedrock downstream of the glacier snout (Broadbent, 1974). The gravels are assumed to be younger than 12 ky. Depths estimated by completing valley wall profiles are shown in table 7. The valley wall profile depth for the upper Tasman Valley (520 m) is in fair agreement with the seismic depth.

Delta advance is determined by a range of techniques. The present position of all lake head deltas is determined by airphotographs. The position at 10 to 20 y ago by earlier airphotographs, at 60 to 110 y ago by cadastral surveys, and at 300 to 1200 y ago by radiocarbon dating. Total postglacial growth is from the area of the river floodplain fill. In order to determine delta growth rates the different positions for each delta were mapped to the same scale and superposed. The delta growth rates for the first three methods were converted to a volume by multiplying the advance by the maximum lake depth and masses calculated using an assumed density of 1.8 t m^{-3} . For the total postglacial rate and to determine the amount of silt in the lake the slopes of the valley sides were taken into account and parabolic sections with area two-thirds maximum fill depth times valley width were used to calculate fill volume.

At the north end of Lake Wanaka the Makarora River delta provides a good example of how the four techniques are applied. Growth appears to have been fairly regular, but it must be noted that the data is incomplete as shown by dashed lines (Fig. 33).

For the bottom sediments all that is known is that at Lake Pukaki a single 0.5 m core of silt, of assumed density 1.25 t m^{-3} , accumulated at 14 mm y^{-1} for the last 16 y from the bomb fallout radiation pattern (J. Irwin, N.Z. Oceanographic Inst., pers comm.). The bottom sediments are layered (Irwin, 1972, fig. 3), and sub-bottom profiling indicated 60+ m of the layered sediment without reaching bedrock (Irwin, pers comm.). The thickness of silt in Lake Pukaki and the other glacial lakes as estimated from extrapolating valley wall profiles is shown in Table 7. For Lake Pukaki the rate for the last 12 ky is 27 mm y^{-1} , twice that for the last 16 years which represents normal events only.

Total lake sedimentation rates (delta plus bottom fill) include abnormal as well as normal events. Eight of the ten rates in table 7 are somewhat larger than the normal rates predicted by the erosion-runoff relation (equation 30), but until better fill depths are determined the exact amount larger is

uncertain. For the other two lakes, Wanaka and Wakatipu, estimated rates are only half the predicted rates and indicate that either the depth estimates are too conservative or equation 30 does not apply to the southern end of the Southern Alps.

On the Canterbury shelf submerged beaches of Otago quartz gravel were deposited when the Canterbury greywacke gravel was being trapped within glacially over-widened valleys. Greywacke gravels form the present beach, the traps having filled and overflowed. The upper Waimakariri, Rakaia and Rangitata valleys have the largest over-widened valleys. Assumed parabolic sections with area two-thirds maximum fill depth times valley width were used to calculate fill volume. The density of fill was taken as 1.8 t m^{-3} . Relevant data and the total mass trapped are given in Table 8. Since the ice retreated 12 ky ago the mass trapped is slightly less than the normal load and a roughly equal amount is assumed to have passed through the traps.

Postglacial fans on the west side of the Tasman Valley, when each is assumed to trap the same proportion of total load, provide a measure of the increase in the rate of erosion towards the crest of the Southern Alps. Data is given in table 9 showing that erosion doubles for each 8 km towards the crest. The fans contain in total 0.6 Gt, and as their catchments are 11% of the Lake Pukaki catchment for which a total trapped mass of 105 Gt has been determined (Table 7), 12 Gt has been eroded from the fan catchments and the fans themselves contain only 1/20 of the total load. Total erosion rates increase from 1.3 to 7.5 mm y^{-1} up the 20 km length of valley.

The extent of erosion into last glaciation aggradation surfaces provides a further check on bedload transport rates. The surfaces are considered here (contrast Lensen, 1968) to be almost synchronous and about 12 ky old. Volumes eroded have been estimated by completing smoothed contours on the remaining surface, with the detail being taken from 1:63 360 topographic sheets, and the average erosion rate calculated using a density of 2.1 t m^{-3} and the 12 ky age. Figure 34 shows the location of the surfaces studied and the rates, and a plot of the rates against calculated bedload rates which shows again that equation 20 gives realistic rates.

Man-made lakes can provide accurate estimates of normal river load, but within New Zealand there are only a few examples and all are discussed below.

In the North Island the Matahina dam on the Rangitaiki River has trapped 0.18 Mt y^{-1} of sediment for the last nine years and a further 30% has passed over the dam (Thompson and Adams, in press), giving a rate of $90 \text{ t km}^{-2} \text{ y}^{-1}$.

At Mangahao near Palmerston North there are two dams, an upper which is continuing to trap gravel and a lower that filled with silt. The lower dam is the only large New Zealand dam to have been sluiced out and it provided a rate of $2000 \text{ t km}^{-2} \text{ y}^{-1}$ (Thompson, 1976).

In the South Island the Benhopai dam was built to supply power to Blenheim in 1927. The lake filled with sediment and about 0.34 Mt y^{-1} or $450 \text{ t km}^{-2} \text{ y}^{-1}$ was trapped before gravel passed over the dam in 1947 (Marlborough Catchment Board drawing 197, 1965). The plant is shut down during floods and after floods the head pool is dug out.

The Clutha River drains an area of Otago the north-western part of which, near the crest of the Southern Alps, is wet, and the south-eastern dry. The Shotover is the only river to flow directly from wet to dry, the others flowing through lakes. Three dams give rates for the area. The Oxburn, near the head of Lake Wakatipu, is in the wet part and traps $17\,000 \text{ m}^3 \text{ y}^{-1}$ of bedload (Otago Central Electric Power Board, pers comm.). From the quartz content of the bedload, the bedload is a tenth the total load, and the total erosion about $10\,000 \text{ t km}^{-2} \text{ y}^{-1}$. The Fraser, in the dry part, gives rates of $41 \text{ t km}^{-2} \text{ y}^{-1}$ (Thompson, 1976). The Roxburgh, by far the largest of the three, drains both wet and dry areas and receives load from the Shotover. It gives rates of $293 \text{ t km}^{-2} \text{ y}^{-1}$ (Thompson, 1976), which from the Oxburn and Fraser rates is thought to be almost all from the Shotover.

The load calculations made in the preceding two chapters are confirmed in part by the spot checks provided by the rough estimates of sediment accumulated in onshore traps. The confirmed load calculations show that about $700 \pm 200 \text{ Mt y}^{-1}$ are eroded from the South Island each year and that the rate is about the same as the tectonic uplift amount. The best way of refining the onshore estimates and so checking the amount in more detail would be to determine fill depths precisely.

Figure 32. Map of South Island showing the sediment traps discussed in the text.

Figure 33. Sketch map showing growth of Makarora River delta at Lake Wanaka. Old delta shorelines are dashed where data incomplete. Beach ridges identified by letters.

Figure 34. Map of central part of South Island showing location of eroded aggradation surfaces and the average annual amount removed from each in kt y^{-1} . Graph on log-log scales shows the good agreement between amount removed and bedload transport rates calculated according to equation 20 and taken from table 2.

Table 6. Table giving data and calculated mass trapped in the overfilled fiords of Westland. Fill depths are assumed (see text), fill density is taken as 1.8 t m^{-3} , and fill age is taken as 12 ky. Erosion rates are calculated for that part of the catchment upstream of the Alpine Fault.

River	Length km	Width m	Depths			Volume 10^9 m^3	Rate Mt y^{-1}	A km^2	Erosion	
			Y m	Z m	Y+Z m				t km y^{-1}	mm y^{-1}
Ahaura	15.2	2400	30	100	130	4.7	0.7	800	890	0.3
Haupiri	7.6	2300	30	100	100	2.3	0.3	88	3860	1.5
Taramakau	28.5	2300	100	230	330	21.6	3.2	994	3250	1.2
Arahura	6.6	650	20	100	120	0.5	0.1	126	610	0.2
Hokitika	20.0	9000	20	230	250	45.0	6.7	793	8510	3.2
Mikonui	6.3	900	30	100	130	0.7	0.1	73	1510	0.6
Kakapotahi	5.5	800	50	100	150	0.7	0.1	36	2750	1.0
Waitaha	11.4	3000	20	100	120	4.1	0.6	151	4075	1.5
Wanganui	22.3	3500	40	230	270	21.0	3.2	337	9380	3.5
Poerua	17.7	1450	50	100	150	3.8	0.6	60	9620	3.6
Whataroa	25.3	6400	50	230	280	45.3	6.8	515	13200	5.0
Waiho	15.2	4270	60	230	290	18.8	2.8	200	14100	5.3
Cook	19.3	4000	60	230	290	22.3	3.4	267	12600	4.7
Karangarua	29.1	3500	20	230	250	25.4	3.8	363	10500	4.0
Makawhio	7.0	2000	20	100	120	1.7	0.3	119	2110	0.8
Mahitahi	7.6	1800	20	100	120	1.6	0.2	155	1600	0.6
Paringa	11.1	1600	20	100	120	2.1	0.3	232	1380	0.5
Moeraki	7.0	800	20	100	120	0.7	0.1	67	1500	0.6
Haast	29.1	1350	25	230	255	10.0	1.5	1324	1134	0.4
Total						232	35			

Table 7. Table giving data and calculated mass trapped in the glacial South Island lakes by different methods P: airphotographs, C: cadastral surveys, R: radiocarbon, T: total postglacial fill, d: delta fill, l: lake fill, t: total fill, and $7.5E6 = 7.5 \times 10^6$. Data references in lower case letters a: photos 1696/31, 3206B/7, b: photos 1698/15, 3206A/1, c: photos 1698/17, 3206A/2, d: photos 2223/2, 3507C/7, e: photos 2714/62, 2715A/29, 3034A/7, f: Christchurch CD 2062, g: photos 826C/16, 3710/19, h: Dunedin CD 8846, CD 2058, i: NZ4231A, j: NZ4232A, k: photo 3709/18, Dunedin CD 964, l: photo 1843E/1, Dunedin CD 8320, m: photo 1843F/1, Dunedin CD 1837, CD 4464. Remainder of data from NZMS1 maps.

Lake (Catchment area km ²)	Site	Age y	Method	Ref.	Fill			Fill cross- section m ²	Volume m ³	Density t m ⁻³	Y		E	
					length m	width m	depth m				kt y ⁻¹	t m ⁻²	y ⁻¹	
ROTOITI (185)	Travers Delta	19.7	Pd	a	4	670	80		2.1E5	1.8	19			
		12 000	Td		5000	450	80	24 000	1.2E8	1.8	18			
	Delta		Tl			1000	100	67 000						
	Mid lake		Tl			560	120	45 000						
	Outlet		Tl			590	140	55 000						
	Mean Tl	12 000			5350			56 000	3.0E8	1.25	31			
	Total T										50		270	
ROTOROA (346)	D'Urville Delta	19.7	Pd	b	3.5	750	145		3.8E5	1.8	35			
	D'Urville Delta	12 000	Td		3700	750	145	73 000	2.7E8	1.8	40			
	Sabine Delta	19.7	Pd	c	4.7	600	145		4.1E5	1.8	37			
	Sabine Delta	12 000	Td		1800	830	145	80 000	1.4E8	1.8	21			
	Deltas		Tl			740	100	50 000						
	Mid lake		Tl			1100	180	132 000						
	Outlet		Tl			1330	190	170 000						
	Mean Tl	12 000			9500			120 000	1.1E9	1.25	120			
	Total T										180		520	
SUMNER (330)	Hurunui Delta	19	Pd	d	21	1550	135		4.4E6	1.8	416			
	Hurunui Delta	12 000	Td		15 000	1070	175	125 000	1.9E9	1.8	281			
	Delta		Tl			700	70	33 000						
	Middle		Tl			1200	180	144 000						
	Outlet		Tl			500	75	25 000						
	Total Tl	12 000	Tl		9500			67 000	6.4E8	1.25	67			
	Total T										348		1050	
TEKAPO (1330)	Red Stag		Td			2220	440	980 000						
	Trig R		Td			2090	420	880 000						
	Trig J		Td			2090	440	920 000						
	Mean (Godley)	12 000	Td		26 600			925 000	2.5E10					
	Mt Gerald (G-M)	12 000	Td		7600			750 000	5.7E9					
	Trig E		Td			600	300	120 000						
	Trig N		Td			2400	140	230 000						
	Mean (Macauley)	12 000	Td		16 800			175 000	2.9E9					
	Total Td	12 000	Td						3.36E10	1.8	5050			
	Godley Delta		Tl			3000	140	280 000						
	Trig Q		Tl			2700	140	251 000						
	Trig K		Tl			3150	210	441 000						
	Ref.		Tl			1400	120	112 000						
	Mean Tl	12 000			15 000			271 000	4.1E9	1.25	423			
	Total T										5500		4130	
PUKAKI (1250)	Tasman Delta		Td			4120	440	1210 000						
	Roundhill		Td			4310	490	1410 000						
	Hoophorn		Td			3000	550	1100 000						
	2200' contour		Td			2660	520	920 000						
	Mean Td	12 000			38 000			1160 000	4.4E10	1.8	6600			
	Delta		Tl			4750	350	1110 000						
	Grid 92		Tl			4750	300	950 000						
	Mean Tl	12 000	Tl		20 000			1030 000	2.1E10	1.25	2140			
	Total T										8740		7000	

Table 8. Table giving data, and calculated sediment mass trapped in the over-widened valleys of the Waimakariri, Rakaia and Rangitata rivers. Fill density is taken as 1.8 t m^{-3} and age as 12 ky.

River	Site	Length km	Width m	Max depth m	Cross section Area m^2	Volume 10^9 m^3	Rate Mt y^{-1}
Waimakariri	1470' contour		1600	170	180 000		
	Sugar Loaf		1840	330	400 000		
	Cora Lynn		2220	240	360 000		
	2250'		1770	210	250 000		
	Mean	43.7			297 000	13.0	2.0
Poulter	Peveril		1200	180	140 000		
	Forest Peak		1010	130	90 000		
	Mean	23.1			115 000	2.7	0.4
Rakaia	McCracken		3170	330	700 000		
	Double Hill		3480	390	1360 000		
	Mean	27.9			1030 000	28.7	4.3
	1800'		3290	330	720 000		
	2200'		1900	240	300 000		
	2800'		920	110	70 000		
	Mean	27.2			360 000	9.9	1.5
Mathias	1900'		2310	330	510 000		
	2100'		1520	180	180 000		
	Mean	16.5			345 000	5.7	0.9
Wilberforce	1800'		2530	220	370 000		
	2200'		1450	180	170 000		
	Mean	22.1			270 000	6.0	0.9
Rangitata	Trig B		2090	270	380 000		
	1500'		3610	200	480 000		
	Craig Phillips		3600	360	860 000		
	Mean	37.7			570 000	21.6	3.2
Clyde	2000'		1770	260	310 000		
	2500'		1200	210	170 000		
	Mean	14.2			240 000	3.4	0.5
Havelock	2000'		1800	270	320 000		
	2500'		1270	190	160 000		
	Mean	18.7			240 000	4.5	0.7
	Grand Total					95.5	14.3

Table 9. Table giving data and calculated erosion rates for seven postglacial fans on the west side of the Tasman Valley. Density is assumed to be 2.1 t m^{-3} and the age of the first three is taken as 8.5 ky as they postdate the Birch Hill glacial advance. The 1:20 ratio of fan volume to total erosion from the text is used to calculate total erosion rates.

FAN	Distance from 5.7 m y^{-1} contour km	Fan Catchment km^2	Fan Area km^2	Fan Height m	Volume 10^6 m^3	Age ky	Rate for Fan $\text{t km}^2 \text{ y}^{-1}$	Erosion mm y^{-1}
Hoop Horn	10	8.3	1.42	70	33.1	8.5	990	7.5
Birch Hill	13	10.1	1.65	70	38.5	8.5	941	7.1
Freds	16	19.6	2.09	59	41.1	8.5	516	3.9
Bush	19	27.6	2.97	56	55.4	12	350	2.6
Twin	24	24.4	2.82	46	43.2	12	310	2.3
Whale	28	24	2.00	64	42.7	12	311	2.3
Jacks	31	28.9	0.98	85	27.7	12	168	1.3
				Total	591 Mt			

CHAPTER SEVEN

THE SOUTHERN ALPS AS A STEADY STATE
MOUNTAIN RANGE

Rapid uplift of a mountain range leads inevitably to rapid erosion and in the long term erosion must equal uplift. The more rapid the uplift, the sooner the equality is achieved. For most of the Southern Alps erosion roughly balances uplift, as it does in some other mountainous areas given in table 10. In order to analyse uplift and erosion a steady state is assumed to exist in the Southern Alps, there being exact agreement between erosion and uplift.

Only mountains with straight slopes and sharp tops are consistent with the steady state; they are 'spiky' mountains whose height represents a balance between uplift and erosion. To the east of the Southern Alps proper the mountains are lower, have flat tops, and are not steady state.

Runoff is the essential factor controlling mountain height and maintaining the steady state. Runoff in the Southern Alps, like the other areas in table 10, is determined by the amount of water intercepted from humid onshore winds, and the amount intercepted is controlled by a negative feedback relation: mountain growth leads to more intercepted water and more runoff, to more erosion, to lower mountains, and hence to less runoff. A state of balance is achieved in the long term (c. 10^5 y), but not in the short term (c. 10^3 y) because of climatic changes.

To model the growth to steady state of the Southern Alps the relations between the height of the mountain range and the amount of water intercepted, and between uplift, runoff, and height have to be established. Mountain height is taken as being the average height of ridge crests measured on lines 1 km apart parallel to the Alpine Fault and to the axis of the Southern Alps and plotted on a cross section normal to the fault (Fig. 35D). Of the water intercepted almost all runs off, and the observed maximum runoff (R_{Max}) is roughly related to the height in metres of the mountain crest along the range by:

$$R_{\text{Max}} = \text{height} / 300 \text{ m y}^{-1} \quad (36)$$

Mountain height should increase with rate of uplift (mm y^{-1}) and decrease with amount of runoff, and for greywacke mountains of the Southern Alps the relation:

$$\text{Height} = 730 \text{ Uplift} / \text{Runoff m} \quad (37)$$

was found by comparing the heights with the uplift pattern given by equation 1 and the runoff pattern given by equation 24. For rocks less resistant than greywacke the heights will be lower according to their degree of resistance. For the less resistant schists near the Alpine Fault the heights are about threequarters those predicted for greywacke mountains, uplift and runoff being unchanged.

The model has the following initial conditions and constraints (Fig. 35). The entire South Island was a plain at sea level and was underlain on the east side of the Alpine Fault by 12 km of greywacke beneath which there was 15 km of less resistant schist. Uplift is according to the pattern shown in figure 10. It is assumed to have started suddenly and to have then continued uniformly. Erosion is assumed to start lowering the ridge crests when the non-steady state mountains reach 2 km, at which height they are assumed to become steady state spiky mountains. Erosion rates for the steady state mountains are calculated from equation 37, erosion and uplift being assumed equal, with allowance being made for the less resistant schist. The maximum runoff is given by equation 36 and the pattern is assumed to be:

$$\begin{aligned} R &= R_{\text{Max}} && \text{for points more than 5 km west of the divide} \\ \text{and } R &= R_{\text{Max}} \exp(-x/18) && \text{for other points} \end{aligned}$$

where x is the distance in kilometres east from the '5 km west of the divide' point and the second equation is modified from equation 24.

The narrowing against the Alpine Fault means that points distant from the fault move closer with time and so their uplift rate is increased, the horizontal and vertical (uplift) components summing as vectors to the 22 mm y^{-1} narrowing vector. The inclination of the narrowing vector increases from nearly zero 100 km from the fault, to 45° at 18 km, and becomes vertical at 6 km.

In terms of the model all points were lower and further from the fault in the past than at present. The model sums the movements in 5000 year steps of uplift, erosion, and narrowing from the initial conditions to give future profiles. The summing is most easily done by computer, and the program used is given in figure 36. The results are presented in two ways (Fig. 37): as a series of profiles showing the growth to steady state, and as a single cross section showing the total amount uplifted. The model predicts that the mountains were originally higher, and the divide closer, to the Alpine Fault, but that the present state was attained after 1.5 My and that the range is now a steady range. with the position of the greywacke-schist boundary being fixed 18 km from the

fault where the inclination of the narrowing vector is 45° . As flat topped mountains are moved towards the fault they become higher, spiky mountains and their place is taken by lower flat topped mountains that are moved in from the east. Thus although the individual mountains change, the mountain range and the type and height of mountain in each part remains the same.

Although the model shows that the Southern Alps quickly become a steady state range and so give no information on when mountain growth started information is provided by the non-steady state mountains in Otago. A single mature surface is assumed to have extended across Otago east from the Alpine Fault, and to have formed the sea level plain used in the model. A nominal age for the surface comes from two places. At Taieri Mouth, south of Dunedin, the surface is 130 m above sea level and near a 7.5 m interglacial bench (McKellar, 1966), giving an uplift of 0.075 mm y^{-1} for the bench and 1.7 My for the surface. At Tekapo the surface is at 2000 m and near tilted lake shorelines for which Wellman inferred uplift of about 0.5 mm y^{-1} , giving a nominal age of 4 My. The nominal age adopted is 3 ± 1 My, but since the nominal age dates the start of uplift the surface may well be older near and younger away from the fault.

The mature surface is best preserved at low altitudes near the coast, and inland it is higher and more dissected. As the mountains have been uplifted rivers between them have downcut apace with the uplift and hillslope erosion has cut back and narrowed the flat mountain summit. The more rapid the uplift, the higher the mountain, the faster the channel downcutting, the wider the valley, and the narrower the flat summit. Eventually traces of the surface disappear from mountains that are about 2 km high, and the straight hillslopes intersect to give a sharp ridge and a spiky mountain. Since the flat topped mountains of Otago are still growing, the region containing them must be eroding more slowly than it is being uplifted. If the Fraser, Lindis, Manuherikia, and Otematata catchments are considered representative, the mean normal erosion rate from tables in chapter 4 is $90 \text{ t km}^{-2} \text{ y}^{-1}$, and the total about twice this or 0.07 mm y^{-1} of rock. The mean uplift is about 0.1 to 0.3 mm y^{-1} from figure 8, so that the erosion is only a fraction of the uplift.

The Japanese rates from table 10 show that the central zone with uplift of 2.2 mm y^{-1} is in steady state, the outer zone of south-west Japan at 1.5 mm y^{-1} is almost in steady state, and that the north-east and inner south-west zones are not in steady state. The uplift rates and the degree to which erosion rate approaches uplift rate suggests that uplift in all zones may have started at the same time, but that in the zones with the slower rates steady state has not yet been achieved. The inner zone where uplift is six times as rapid as erosion

is described by Yoshikawa (1974) as a highland region dissected by narrow gorges that separate highlands of low relief that are not being so intensely eroded. The gorges are presumably downcutting apace with uplift, but the highlands, eroding more slowly than the uplift, are non steady state flat topped mountains like those in Otago.

The areas of the South Island with each type of mountain are shown in figure 38. When compared to Wellman's uplift map (Fig. 8) the boundary between the two types of mountain corresponds approximately to the 0.4 mm y^{-1} uplift contour. At lower uplift rates the mountains are not in steady state and erosion is not in balance with uplift. Approximate erosion and uplift rates for the areas that are not in balance are:

region	Erosion		Uplift mm y^{-1}
	$\text{t km}^{-2} \text{ y}^{-1}$	mm y^{-1}	
North West Nelson	2360	0.9	1.0-1.5
South Canterbury	500	0.2	0.1-0.4
East Otago	180	0.07	0.1-0.3
Fiordland	550	0.2	0.2-0.3

The areas that are not in balance amount to about 50% of the South Island. The remainder, which as might be expected includes all the areas of rapid uplift, is in balance, and for the South Island as a whole erosion roughly equals uplift. The plate collision zone is thus in a steady state of erosional equilibrium that can continue indefinitely with the upthrusting of the Pacific plate being balanced by equal erosion of the plate edge, the Southern Alps.

Offshore sediments provide the final check for the model and erosion rates inferred from onshore data. Unfortunately no systematic study of deposition rates off the New Zealand coast has yet been made, and much key data is lacking. In particular the ages and thicknesses of deep reflectors shown in seismic surveys are uncertain, and an average seismic velocity in the sediments of 2.2 km s^{-1} had to be assumed. The following is an attempt to synthesise the available data and to determine accumulation rates for the areas shown in figure 39. Both published and unpublished data has been used, the most recent publication being that of Davey (1977).

Off south Westland Norris (1977) has described thick postglacial sediments that blanket the offshore $12 \pm 2 \text{ ky}$ terminal moraines mentioned in the last chapter. The thickness of the postglacial sediments is about 140 m a few

kilometres off the coast and thence decreases away from the coast. Further offshore, airgun seismic surveys made by the Mobil Corporation in 1972 (copies held by the N.Z. Oceanographic Inst.) indicate a clastic wedge of undated, undeformed sediments above a reflector that is probably the Oligocene Cobden Limestone that is about 3 km thick near the coast and that like the inshore sediment also thins offshore halving in thickness for each 45 km. From both sets of data the following deposition pattern is inferred with the inshore rate being $12 \pm 2 \text{ mm y}^{-1}$:

$$\text{rate} = (12 \pm 2) \exp(-0.016 \cdot \text{distance}) \text{ mm y}^{-1}$$

For a 300 km long, 150 km wide strip off south Westland between Cascade Point and Punakaiki integrating the relationship gives $200 \pm 30 \times 10^6 \text{ m}^3 \text{ y}^{-1}$, or $250 \pm 40 \text{ Mt y}^{-1}$ at a density of 1.25 t m^{-3} .

In the deep Tasman Basin Cochran and Osmond (1976) determined deposition rates of 0.044 to 0.0018 mm y^{-1} for terrigenous sediments 600 to 1200 km from the inferred source in south Westland. Their data is consistent with an exponential decrease with distance, but halving every 140 km, more slowly than the inshore rates:

$$\text{rate} = (1.1 \pm 0.2) \exp(-0.0053 \cdot \text{distance}) \text{ mm y}^{-1}$$

The average rate for the 1200 km extent of sedimentation is $0.17 \pm 0.03 \text{ mm y}^{-1}$, but the full 1200 km extent includes 150 km of the shelf already calculated above, and this part must therefore be subtracted. Taking the Tasman Basin sediments as part of a conical ring of exponential cross section (see inset, figure 39) centred on south Westland with radii of 150 and 1200 km, an angle of 100° between the Lord Howe Rise in the north and Fiordland in the south, and with a height of $1.1 \pm 0.2 \text{ mm y}^{-1}$ at source, $0.50 \pm 0.09 \text{ mm y}^{-1}$ at 150 km and zero at 1200 km, and approximating the exponential conical sections to linear sections of equal area, the volume is:

$$\begin{aligned} V &= \frac{100}{360} \cdot \frac{\pi}{3} \cdot ((1200 \times 10^3)^2 (0.34 \pm 0.06 \times 10^{-3}) - (150 \times 10^3)^2 (1.60 \pm 0.3 \times 10^{-3})) \\ &= 132 \pm 27 \times 10^6 \text{ m}^3 \text{ y}^{-1} \text{ or } 165 \pm 35 \text{ Mt y}^{-1} \text{ at a density of } 1.25 \text{ t m}^{-3}. \end{aligned}$$

Off West Nelson between Punakaiki and Cape Farewell only a few metres of silt has been deposited in the last 10 to 20 ky (Van der Linden and Norris, 1974). The silt overlies a prograded clastic wedge like that in south Westland, that is about 0.6 km thick and inferred to be younger than $2.5 \pm 0.5 \text{ My}$. For a 200 km

long, 150 km wide part of the shelf and slope some $1.8 \times 10^{13} \text{ m}^3$ has accumulated at $7.5 \pm 1.5 \times 10^6 \text{ m}^3 \text{ y}^{-1}$ or $9 \pm 2 \text{ Mt y}^{-1}$ using a density of 1.25 t m^{-3} .

Sands that are presently accumulating in the South Taranaki Bight and in Farewell Spit have been moved north from their source in south Westland by longshore drift. Furkert in 1947 made the first estimate of amount. From observations at Westport and on the spit he concluded that some $3.6 \times 10^6 \text{ m}^3$ (6.5 Mt at an assumed density of 1.8 t m^{-3}) of sand came from the Grey and rivers to the south each year, and that a like amount of identical sand was added to the spit each year. From the rates he inferred erosion of about 100 Mt y^{-1} in south Westland, with most of the rock being ground up and washed offshore and only a small part accumulating on the spit. Furkert's work was well ahead of its time, and his estimates were reasonably good. More recently Van der Linden (1969) inferred from seismic profiles from the South Taranaki Bight that 10 to 20 m of silty sand had accumulated over $12\,500 \text{ km}^2$ in the last 19 ky; equivalent to $10 \times 10^6 \text{ m}^3 \text{ y}^{-1}$ or 12.5 Mt y^{-1} , and in addition to the 6.5 Mt y^{-1} added to Farewell Spit. Of the total 19 Mt y^{-1} , some 6 Mt y^{-1} (author's unpublished results) is from the North Island rivers and the remaining 13 Mt y^{-1} is from the South Island.

Off the east coast of the North Island from Napier to Castlepoint Lewis (1973) inferred deposition of 7 Mt y^{-1} on the continental shelf and slope, but most of this is probably from the North Island. Further offshore seismic reflection profiles made by Mobil and figured by Katz (1974) are interpreted here to indicate $170 \times 10^{12} \text{ m}^3$ of Plio-Pleistocene sediments in the Hikurangi Trench and another $20 \times 10^6 \text{ m}^3$ in the southern extension of the Kermadec Trench. Much of the sediment is from Kaikoura and Canterbury and has been carried north by the Southland Current and thence down the Pegasus and Kaikoura canyons into the trench. Taking $2.5 \pm 0.5 \text{ My}$ for the age of the oldest sediments, 90 ± 20 and $10 \pm 3 \text{ Mt y}^{-1}$ have been deposited in each trench.

In Pegasus Bay off Christchurch Herzer (1977) described an accumulation of silt that contains $1 \times 10^{11} \text{ m}^3$ and that from eustatic sea level changes is less than $10 \pm 2 \text{ ky}$ old. At a density of 1.25 t m^{-3} the silt has accumulated at $13 \pm 3 \text{ Mt y}^{-1}$. Further offshore young, undeformed sediments are 1.7 km thick at the 100 m isobath and thin offshore thus (observations from the same Mobil survey discussed above):

$$\text{thickness} = 1.7 \exp(-0.013 \cdot \text{distance}) \text{ km}$$

For a 250 km long, 100 km wide belt parallel to and offshore of the 100 m isobath the integrated volume is $2.4 \times 10^{13} \text{ m}^3$, or $3 \times 10^{13} \text{ t}$ at a density of 1.25 t m^{-3} .

The sediments are inferred to be 2.5 ± 0.5 My old and to represent deposition of $12 \pm 2 \text{ Mt y}^{-1}$. Some of the sediment underlying the Canterbury Bight inshore of the 100 m isobath must also be younger than 2.5 ± 0.5 My. If sediments of this age are 0.7 km thick onshore (chapter 6), about 1.1 km thick from the Resolution I offshore well (NZ Geological Survey open file report PR648), and 1.7 km thick at the 100 m isobath the average thickness will be 1.2 km for a shelf 250 km long and 60 km wide; $1.8 \times 10^{13} \text{ m}^3$ or $9 \pm 2 \text{ Mt y}^{-1}$ at a density of 1.25 t m^{-3} .

Off Otago silty sand from the Clutha River is presently deposited just north of the Otago Peninsula, but during low sea levels probably accumulated on the Otago shelf or in the Bounty Trough. In the Snares Depression off south Otago about $3 \times 10^{13} \text{ m}^3$ have accumulated above reflector 'X' of Davey (1977, fig. 16), and I infer rates of $4 \pm 4 \text{ Mt y}^{-1}$.

Off Southland I interpret two seismic profiles from the Solander Trough figured by Katz (1974) to indicate $6 \times 10^{13} \text{ m}^3$ deposited over 20 ± 10 My or about $4 \pm 2 \text{ Mt y}^{-1}$.

The adopted total sedimentation off the South Island is $580 \pm 110 \text{ Mt y}^{-1}$ - $440 \pm 80 \text{ Mt y}^{-1}$ off the west and $140 \pm 30 \text{ Mt y}^{-1}$ off the east coast - but the error might be twice as large as indicated. Despite the large error, the rate provides substantial confirmation of the onshore sedimentation rates and river load estimates, and of the steady state model for the Southern Alps.

Figure 35. Diagrams normal to the Alpine Fault and on the same horizontal scale showing initial and final constraints for the model discussed in the text. A: Uplift curve taken from equation 1. B: Runoff pattern in metres when mountain divide is 2.25 km high and 20 km from the fault. Actual pattern is related to divide height and position. C: Initial geologic section taken from figure 10. Vectors of 22 mm y^{-1} show direction of movement of eastern side, western side being assumed fixed. No attempt is made to explain the narrowing geologically. D: cross section showing present height of the Southern Alps determined from the average height of ridge crests on lines parallel to the Alpine Fault and normal to the section.

```

                                B U I L D M O U N T A I N S
                                * * * * *

BEGIN

  FILE FCR (KIND=READER), FLP (KIND=PRINTER);

  INTEGER I,J,COUNTER,
           SITPEAK,
           N,
           TIME;
  REAL PEAK,
        RUNOFF,
        KONST,
        REAL ARRAY;
  D,
  E,
  H,
  MH,
  P,
  R,
  RESIST,
  SCHIST,
  TE,
  THETA,
  TP,
  TR,
  TUT,
  U,
  UT,
  V(126);

  % COUNTERS
  % POSITION OF CREST
  % NUMBER OF POINTS
  % TIME IN UNITS OF 5000 YEARS
  % HEIGHT OF DIVIDE
  % MAXIMUM RUNOFF
  % RADIAN TO DEGREE CONVERSION
  % DISTANCE
  % EROSION IN MM PER YEAR
  % HEIGHT OF POINT IN METRES
  % MEAN HEIGHT OF POINT
  % POSITION RELATIVE TO FAULT
  % RUNOFF AT POINT
  % RESISTANCE OF ROCK
  % LITHOLOGY AT POINT
  % EROSION*100
  % INCLINATION OF NARROWING
  % POSITIONS*10
  % RUNOFF*100
  % TOTAL UPLIFT /10 IN METRES
  % UPLIFT
  % TOTAL UPLIFT
  % VELOCITY OF NARROWING

  READ(FCR,<I2>,N);
  READ(FCR,<I2,25I3>,>D);
  KONST:= 180/(4*ARCTAN(1));
  WRITE(FLP,< 26I5>,>D);
  FOR I:=1 STEP 1 UNTIL N DO
    BEGIN
      UT(I):= SCHIST(I):= H(I):= 0;
      R(I):= 0(I);
    END;
  FOR TIME:=1 STEP 1 UNTIL 650 DO
    BEGIN
      COUNTER:=+ 1;
      PEAK:=0;
      FOR J:=1 STEP 1 UNTIL N DO
        BEGIN
          IF P(J) GTR 6 THEN V(J):= SORT(468=U(J)*U(J))
            ELSE V(J):=0;
          P(J):= * - V(J)/200;
          U(J):=12.892* (P(J)+1)**0.52456 *EXP(-0.07204*(P(J)+1));
          H(J):= * + U(J)*5;
          UT(J):= * + U(J)*5;
          RESIST(J):= IF UT(J) LSS 12000+H(J) THEN 1.0
            ELSE IF UT(J) LSS 27000+H(J) THEN
              1.2-0.0000166*(UT(J)-H(J))
            ELSE 0.75;
          IF V(J) NEQ 0 THEN THETA(J):= ARCTAN(U(J)/V(J)) * KONST
            ELSE THETA(J):=90;
          IF (H(J) GTR 1.04*PEAK) THEN
            BEGIN
              PEAK:= H(J);
              SITPEAK:=J;
            END;
        END;
      RUNOFF:= PEAK/300;
      FOR J:=1 STEP 1 UNTIL N DO
        BEGIN
          IF P(J) LEQ P(SITPEAK) = 5
            THEN R(J):= RUNOFF
            ELSE R(J):=RUNOFF*EXP(-(P(J)-P(SITPEAK)+5)/18);
          E(J):= IF UT(J) GTR 2000 THEN H(J)*R(J)/(RESIST(J)*730)
            ELSE 0;
          H(J):= * - E(J)*5;
          MH(J):=H(J) + E(J) *2.5;
          IF UT(J) GTR (12000 + H(J)) THEN SCHIST(J):=1;
          TP(J):=P(J)*10;
          TR(J):=R(J)*100;
          TE(J):=E(J)*100;
          TUT(J):=UT(J)/10;
        END;
      IF COUNTER EQ 10 THEN
        BEGIN
          COUNTER:=0;
          WRITE(FLP,<///,3R10,2,1R,/>,>PEAK,P(SITPEAK),RUNOFF,TIME*5000);
          WRITE(FLP,< 26I5>,>SCHIST,TR,TE,MH,TUT,TP,THETA);
        END;
      END;
    END;
  END;

```

Figure 36. Computer program in Burroughs B6700 Extended Algol for calculating the growth to steady state of the Southern Alps according to the proposed model.

Figure 37. Top. Five profiles normal to the Alpine Fault showing growth to steady state of the Southern Alps after 0.1, 0.3, 0.6, 1.0, and 2.5 My. Vertical exaggeration = 6.7 times. Bottom. Sketch to equal vertical and horizontal scales showing amount eroded and amount remaining after 2.5 My of uplift.

Figure 38. Map of South Island showing the areas with spiky mountains and the areas with flat topped, non steady state mountains.

Figure 39. Map showing offshore areas in which the sediment eroded from the South Island is deposited. Rates in each area are given in Mt y^{-1} .

Table 10. Table giving average uplift and average erosion rates for six areas of the world where uplift and erosion are roughly equal. A New Zealand rate from an area of comparable size, the entire South Island, has been calculated from the point values in the text.

Area	Source	Erosion		Uplift
		$t \text{ km}^{-2} \text{ y}^{-1}$	mm y^{-1}	
Taiwan	Li 1976	13650	5.2	5.5
European Alps	Clark & Jager 1969	1000-1900	0.4-0.7	0.4-1.1
Japan:	Yoshikawa 1974			
Central		5860	2.2	2.2
Outer SW		3380	1.3	1.5
North east		850	0.32	1.3
Inner SW		270	0.10	0.6
Average		2260	0.85	1.3
New Guinea	Pain & Bowler 1973		0.8-1.0	?
	Simonett 1967		1.0-2.2	?
	after Smith 1970		?	$\approx 2-3$
Himalayas	Curray & Moore 1971		0.7	?
South Island, NZ				
'average'	this thesis	4000 $\pm$ 1000	1.5 $\pm$ 0.4	1.5 $\pm$ 0.1
maximum	this thesis	>27 000	>10	22 $\pm$ 2

CHAPTER EIGHT

CONCLUSION

Uplift and erosion are roughly in balance in the Southern Alps, and the following rates have been determined:

tectonic uplift	$620 \pm 20 \text{ Mt y}^{-1}$
river load	$700 \pm 200 \text{ Mt y}^{-1}$
offshore deposition	$580 \pm 110 \text{ Mt y}^{-1}$

From the rate data and from geological constraints a model for the growth of the Southern Alps has been inferred which demonstrates how the mountain range reached and maintained steady state.

The Southern Alps, being athwart the roaring forties are a special mountain range and simpler than most, but other mountain ranges could be treated in the same way and many are expected to be in steady state.

In New Zealand the uplift rates have been established over the last ten years, but the river load and deposition rates are only now being investigated and the estimates made here might be substantially refined. Almost all topics in the thesis need further detailed work, but the most productive work will probably be determining fill depths in the glacial lakes and fiords and estimating landslipping caused by the largest earthquakes.

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Appendix 1. Table giving erosion rates for 16 parts of the South Island taken from the literature. In the reference column a '*' indicates rates taken from data in the cited source, and the source is indicated by the lower case letters:

a: Adams, unpublished work. b: Brown, L. J. 1972, N.Z. Geological Survey report 60.
 c: Pemberton, G. 1974 unpublished report to Nelson Catchment Board. d: Benson, W. N. 1946 N.Z. Jour. Sci. and Tech. 27B: 420-426. e: O'Loughlin, C. L. 1969 N.Z. Jour. Geol. Geophys. 12: 684-706.
 f: Hayward, J. A.; Blakeley, R. J. 1976 N.Z. Hydrological Soc. Symp. Rotorua. g: Howard, R. 1970 Soil and Water 6(3/4): 2-4. h: Speight, R. 1941 Trans. Roy. Soc. N.Z. 71: 128-133. i: Cuff, J. 1974 Report to South Canterbury Catchment Board. j: Files in Christchurch District Office, Ministry of Works and Development. k: Marshall, P. 1912: Trans N.Z. Inst. 45: 342-343. l: Leamy, M. L. 1969 N.Z. Jour. Sci. 12: 373-379. m: MWD Head Office File 92/11/90/2/2. n: Thompson, S. M. MWD Power Division pers. comm.

Area	Place	Method	Reference	rates ($t\ km^{-2}\ y^{-1}$)		
				bedload	suspended load	total load
Picton	Barnes Dam	Dam siltation	a*			115
Blenheim	Wairau lagoon	Lagoon siltation	b*		260+	
Nelson	Waimea R.	Gravel extraction	c	134		
Murchison	L. Matiri	Lake siltation	a			100+
Clarence	Wharekuri Str	Suspended load sample	a		2500	
Christchurch	Waimakariri R.	Suspended load samples	d		770	
Christchurch	Centre Crk.	Channel degradation	e*	170		
Christchurch	Torlesse Str.	Bedload trap	f	330		
Westland	Waitangi R.	Channel aggradation	g	2250		
Westland	Franz Josef	Lake siltation	h*			4700+
South Canterbury	Opihi	Erosion assessment	i*	77+		
Tasman Valley	Devils Elbow	Sediment trap	j	350		
Tasman Valley	Hooker R.	Suspended load samples	k		1070	
Hawea	Hawea Flat	Aggrading fans	l*			1800
Hawea	Hunter R.	Delta siltation	m*		1000+	
Southland	Mararoa R.	Channel surveys	n	63		