15

Rates of uplift and denudation

15.1 Tempo of geomorphic change

Having examined the operation of the various endogenic and exogenic geomorphic processes in Parts II and III we can now look at the way these two sets of processes interact in the formation of the landscape. It is appropriate to begin with the overall rates at which endogenic and exogenic processes operate since it is the balance between uplift and denudation at any particular point on the Earth's surface that determines whether that point becomes higher or lower with respect to sea level. Through our discussion of endogenic and exogenic progresses it will have become apparent that the rates at which they operate vary enormously; in fact a glance at Figure 15.1 shows that their rates range over some 16 orders of magnitude.

The rates of operation of geomorphic processes may be expressed in various ways. One approach is to view the transport of materials in terms of geomorphic work and to represent this in units of power. This can be done by relating the operation of a particular process to its duration. Although this approach is useful for comparing the relative importance of specific geomorphic processes, when considering the overall rates at which the landscape changes it is more usual to express this in terms of an average lowering or raising of the ground surface over a period of time. Since we are concerned in this chapter primarily with the net effect of various geomorphic processes in modifying the landscape this is the approach we will use here.

Changes in the average elevation of the ground surface are most commonly expressed in terms of mm ka^{-1}. For long periods of time of the scale of millions of years it is more appropriate to use the equivalent unit of m Ma^{-1} (note that 1 mm ka^{-1} equals 1 m Ma^{-1}). Very rapid processes may be better described in terms of mm a^{-1} or even m s^{-1}. It has become fashionable in recent years to express changes in ground elevation in 'Bubnoff units', often represented as B. This is simply equivalent to a change in ground elevation expressed in mm ka^{-1} or m Ma^{-1} and seems to represent a rather needless proliferation of units.

Since rates of denudation are frequently estimated from the *mass* of sediment transported by rivers, glaciers or the wind, we need to be able to convert this mass to a *volume* in order to estimate the equivalent rate of ground lowering. This is done by relating the mass of material transported to its mean density, usually assumed to be the density of typical crustal rocks (approximately 2700 kg m^{-3}). Such a procedure, however, can raise difficulties when comparing the removal of solid and dissolved constituents by rivers. This problem is examined in Section 15.3.1.

15.2 Rates of uplift

Before we discuss how rates of uplift can be measured it is important to clarify what we actually mean when we use the term 'uplift'. This is necessary because it is used in two quite distinct senses. Here we will use the term **surface uplift** to refer to the upward movement of the landsurface with respect to a specific datum (normally sea level). What we will term **crustal uplift**, on the other hand, refers to the upward movement of the rock column with respect to a specific datum (also normally sea level).

As is apparent from Figure 15.2 the relationship between surface uplift and crustal uplift depends on the prevailing rate of denudation. Crustal uplift will only equal surface uplift if there is no denudation during the period of uplift. If, as is likely, denudation does occur then crustal uplift will be greater than surface uplift. But if crustal uplift is exactly matched by denudation then there will be no surface uplift. In fact crustal uplift is frequently exceeded by denudation, and in this situation surface elevation is obviously reduced.

In most cases surface uplift is associated with active tec-

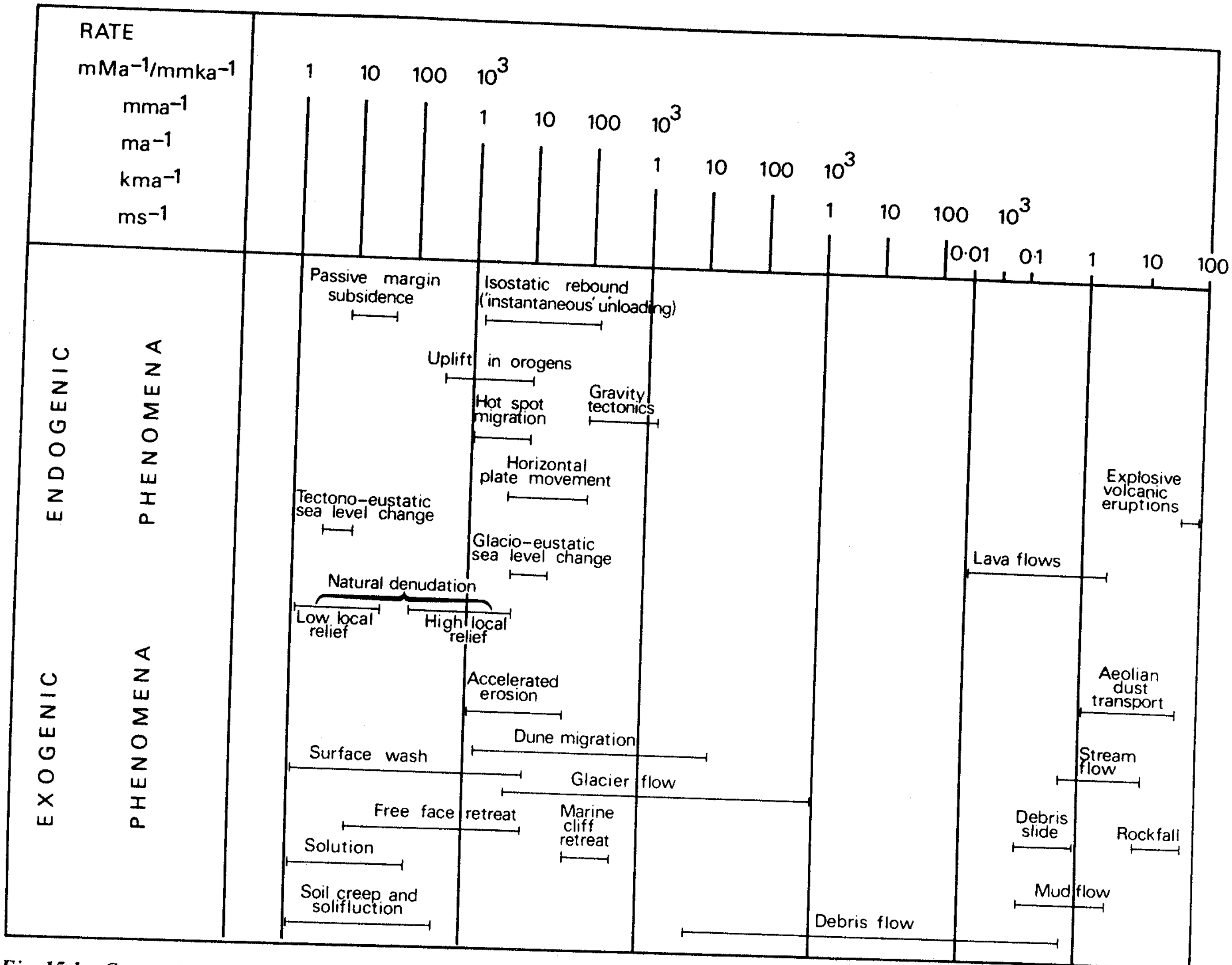

Fig. 15.1 Comparison of rates of various endogenic and exogenic geomorphic processes. Note that rates of specific slope processes are discussed in Chapter 7 and are included here for comparative purposes.

tonic processes, but crustal uplift can occur simply as the inevitable isostatic response to denudation; that is, the removal of material through denudation reduces the load on the crust and it moves upward to restore isostatic equilibrium. In general such crustal uplift resulting from isostatic rebound does not exceed the rate of denudation so there is no net surface uplift; nevertheless, where flexural effects are important (see Section 4.2.3), or where a landsurface is locally dissected by deep gorges, such isostatic recovery can lead to surface uplift of parts of a landscape.

15.2.1 Methods of measurement and estimation

Rates of uplift can be determined either directly or indirectly by a range of techniques appropriate to different time scales (Fig. 15.3). Some of these methods measure or estimate surface uplift, whereas others indicate rates of crustal uplift, and it is important to distinguish between them.

Geodesy is concerned with the precise determination of the Earth's surface form and **geodetic** methods can provide information on short-term rates of surface uplift. One way this can be done is through precise levelling surveys which are repeated after a number of years or, more often, several decades. Although systematic errors in levelling make it difficult in some cases to distinguish a real vertical movement from an artefact of the measuring procedure, such geodetic techniques provide an important basis for documenting surface uplift and subsidence over periods of a few years to a few decades. The technology is now available whereby satellites can determine with great accuracy the form of the Earth's surface (**satellite altimetry**) and changes in the geoid over time have already been recorded by this means.

Along coasts tide gauge records can provide high-quality information on vertical movements of the landsurface, as long as allowance is made for the slow, continuing, post-

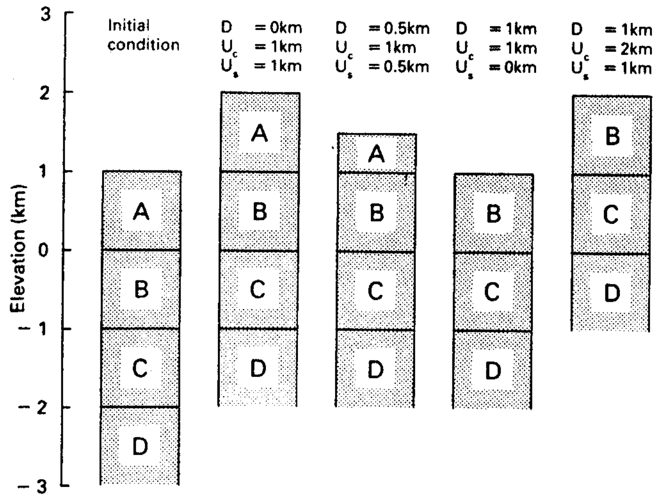

Fig. 15.2 *Schematic illustration of the distinction between surface uplift and crustal uplift.* D *designates the depth of denudation and* U_c *and* U_s, *respectively, the amount of crustal and surface uplift. Note that if local (Airy) isostasy is to be maintained then denudation of a given amount will be accompanied by crustal uplift of around 80 per cent of the depth of denudation (see Figure 4.7.)*

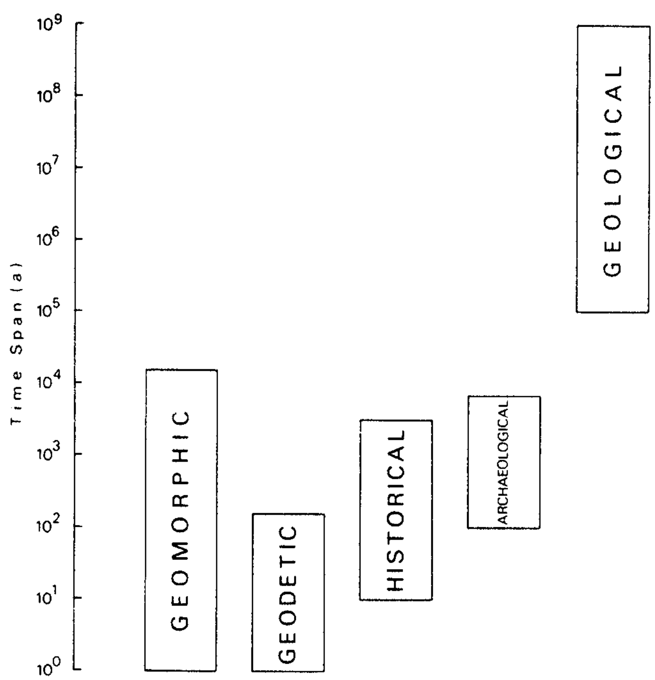

Fig. 15.3 *Appropriate time scales for different types of evidence used to determine rates of uplift.*

glacial sea level rise (currently about 1 mm a^{-1}). Tide gauge stations are relatively numerous in some areas of the world. In a few instances their records extend back into the last century, but the quality of some of these early records is too poor for their use in precise calculations of surface uplift or subsidence rates.

As the time period with which we are concerned becomes longer, so the means of determining rates of vertical movement become more indirect and less precise. Historical and archaeological evidence can be valuable in certain circumstances. Surface uplift in Iraq around AD 100–200 has been inferred from the breaching of an irrigation canal, and a similar instance has been reported of the disruption of an Inca canal in Peru. In the coastal regions of Greece, a tectonically active area, rapid vertical movements have been documented from the displacement of buildings and other constructions built in antiquity. Discrete phases of surface uplift and subsidence have been inferred from the displacement of the northern end of the ramp (*didkos*) used by the Greeks to haul ships across the Isthmus of Corinth.

Geomorphic evidence of vertical movements is represented by dated landform features which can be related to a previous absolute or relative elevation. Such evidence might include dated basalt flows displaced by faulting or elevated wave-cut platforms or beaches which can be dated from their associated deposits. One of the best-documented uplift histories in the world is to be found in the marine terraces of the Huon Peninsula, Papua New Guinea. Here a series of coral terraces has been cut by high sea levels during the Late Pleistocene as the coastline has risen rapidly. The rates of uplift can be established because the age of each coral terrace can be determined by radiometric dating (see Sections 15.2.2 and 17.2.1).

Geological evidence for uplift is usually relevant to periods of millions of years and is established from the elevation of rocks of known age formed at, or below, sea level, and from a variety of dating techniques which enable estimates to be made of the depth of burial of a rock at a known time in the past. The present-day height of marine deposits above present sea level can be used to calculate a long-term mean rate of surface uplift if the sea level at the time of the formation of a deposit is known. Such evidence, however, only provides an estimate of net surface uplift since there could have been intervening periods of subsidence.

Increasing use is now being made of radiometric dating to estimate long-term rates of rock uplift relative to the landsurface (Fig. 15.4(A)). The 'clocks' of radiometric dating procedures, such as K–Ar and Rb–Sr dating, are only set once the rock has cooled below a certain temperature, known as the **closure temperature**; this is the point at which the system becomes closed and the products of radioactive decay are retained (see Appendix B). If the geothermal gradient (typically 20–30 °C km^{-1} in the upper crust) is known, the depth below the surface at which the radiometric 'clock' started can be calculated. The radiometric age of the rock then gives the time taken for denudation to remove the overlying strata and expose the dated rock at the surface. This information provides us with a mean denudation rate, and if we assume that there has been no change in surface elevation then this will be equal to the

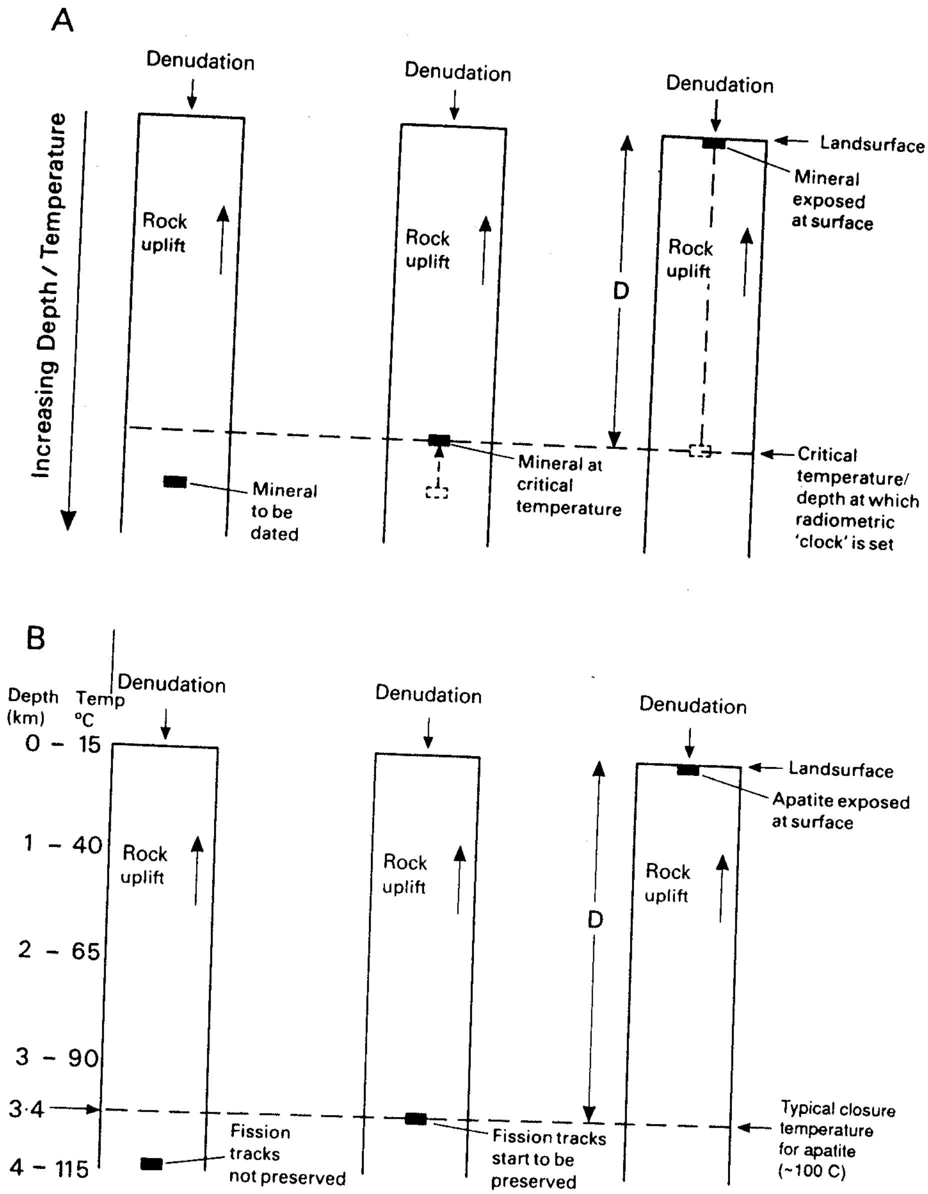

Fig. 15.4 Schematic illustration of the use of radiometric dating (A) and fission track dating (B) in the estimation of denudation and uplift rates relative to the landsurface. In (A) the age of a mineral at the surface is determined by the time elapsed since it passed through the critical isotherm at which its radiometric 'clock' was set. This isotherm, and therefore depth, varies for different minerals. The rate of denudation is calculated by dividing depth (D) by age. In (B) the estimation of denudation using fission track dating of apatite is illustrated. If we assume that the geothermal gradient is 25 °C km^{-1} (as indicated) and the age of apatites exposed at the surface is 10 Ma then the mean denudation rate is 3400 (m)/10 = 340 m Ma^{-1}.

mean crustal uplift rate relative to sea level. Note that, used alone, this technique does not provide any direct information on the amount of absolute uplift that has occurred.

The estimation of crustal uplift rates using radiometric dating is most appropriately applied to orogenic belts since it is here that metamorphic events associated with significant heating of the crust are most likely to 'reset' the radiometric clock by overprinting the original radiometric dates established when the rocks were first formed. This enables a rock uplift rate relative to the landsurface to be estimated for a short period of a few million years since the resetting event. It is also only in active orogenic belts that denudation and crustal uplift rates are likely to be approximately equal for a significant period of time.

An elaboration of this technique is to compare the radiometric ages of two different minerals which start their radiometric clocks at different temperatures. The radiometric clock for the mineral muscovite, for instance, starts about 200 °C higher than the radiometric clock for biotite. If the geothermal gradient is 20 °C km^{-1} this is eqivalent to a difference in depth of 10 km, so the difference in age between the two minerals when exposed together at the surface represents the time taken for 10 km of denudation to occur. Again, assuming a rough equilibrium between crustal uplift and denudation rates, this provides an estimate of the former.

Another technique based on a similar principle is **fission track dating** (Fig. 15.4(B)). The fission of uranium-238, which is relatively abundant in minerals such as apatite found in granites and other basement rocks, involves the break up of the original nucleus into two fragments of roughly equal mass. These charged particles recoil from each other and move in opposite directions thereby creating microscopic dislocations or 'tracks' in the crystal lattice. The number of tracks formed is a function of the concentration of uranium present and the time elapsed since tracks began to be preserved. Determining the uranium concentration and counting the number of microscopic tracks formed enables an age to be estimated. The particular advantage of the technique for estimating rock uplift relative to the land-surface and denudation rates is that for the mineral apatite the fission tracks are destroyed, or **annealed**, at temperatures above about 100±20 °C. The fission track closure temperature for apatite is, therefore, located at a relatively shallow depth in the crust corresponding to this temperature. Depending on the geothermal gradient, this **fission track annealing zone** for apatite usually lies at a depth of between 3.5 and 5.5 km.

As with the radiometric methods already discussed, this technique can only provide estimates of crustal uplift rates indirectly through the calculation of denudation rates. Outside active orogenic belts where these two rates will rarely be similar, the main use of both radiometric and fission track techniques is in the estimation of denudation rates (see Section 15.4).

15.2.2 Spatial and temporal variations

From our discussion of the mechanisms of orogeny and epeirogeny in Chapters 3 and 4 we would expect considerable differences in their associated rates of uplift. Although this is clearly the case in the long term, evidence from relevelling surveys provides some equivocal data on rates of uplift in regions affected by epeirogenic movements.

15.2.2.1 Orogenic uplift

In active orogenic zones significant vertical movements of the surface up to several metres may occur in a single earthquake. In the major Alaskan earthquake of 1964, for example, an area of some 400 000 km^2 was raised by an average of 2 m and a maximum of 12 m. Although there may be rapid variations in rates of vertical displacement in space and time along convergent and transform plate margins, high mean rates of uplift can be sustained for periods of hundreds of thousands of years (Table 15.1). An extreme example is the Ventura Anticline at the western margin of the Transverse Ranges in southern California which has experienced crustal uplift at an average rate of at least 16 000 m Ma^{-1} over the past 1 Ma. Rather more modest rates of surface uplift of up to 3300 m Ma^{-1} have been recorded from the elevated coral terraces of the Huon Peninsula (Fig. 15.5).

Minimum crustal uplift rates in major orogenic belts, such as the Alps and the Himalayas, located at convergent plate margins range from 300 to about 800 m Ma^{-1} averaged over periods of several million years (Table 15.1). But rates may be much higher than this; in southern Tibet, for

Table 15.1 Long-term mean uplift rates in orogenic zones determined by various methods

LOCATION	METHOD	RATE (m Ma^{-1})	PERIOD	SOURCE
Central Alps	Apatite fission track age	300–600*	6–10 Ma BP	Schaer *et al.* (1975) *Tectonophysics* **29**, 293–300
Central Alps	Rb and K–Ar apparent ages of biotite	400–1000*	10–35 Ma BP	Clark and Jäger (1969) *American Journal of Science* **267**, 1143–60
Kulu–Mandi Belt, Himalayas	Apparent Rb-Sr ages of coexisting biotites and muscovites	700*	25 Ma BP–Present	Mehta(1980) *Tectonophysics* **62**, 205–17
Southern Alps, New Zealand	Apparent K–Ar ages of schists	10 000*	1 Ma BP–Present	Adams (1981) *Geol. Soc. Lond. Spec. Publ.* **9**, 211–22
Southern Alps, New Zealand	Estimated ages of elevated marine terraces	5000–8000†	140 ka BP–Present	Bull and Cooper (1986) *Science* **234**, 1225–8
Huon Peninsula, Papua New Guinea	U-series and ^{14}C dating of elevated marine terraces	1000–3000†	120 ka BP–Present	Chappell (1974) *J. Geophys Res.* **79**, 456–64

* Crustal uplift.
† Surface uplift.

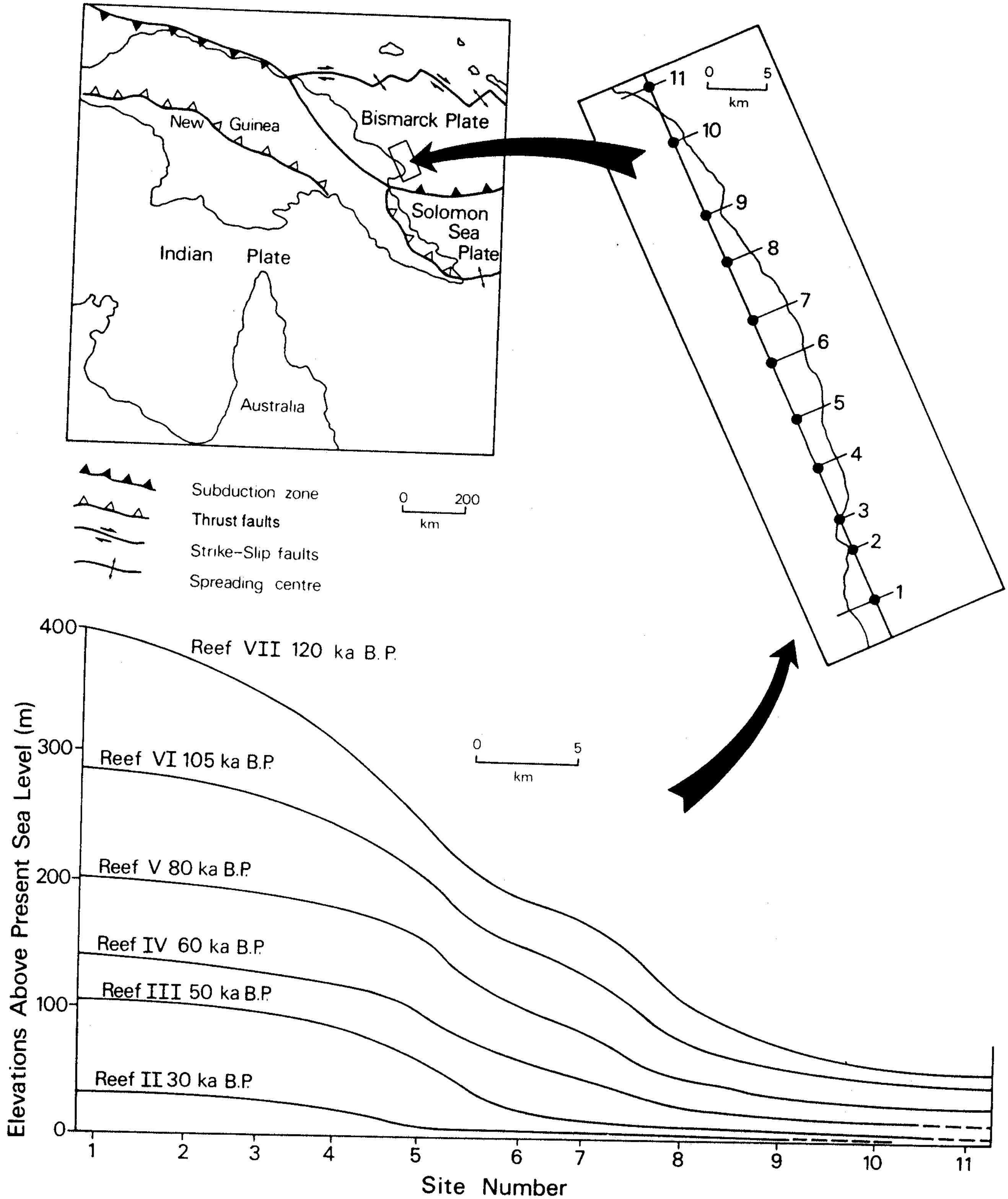

Fig. 15.5 Tectonic setting and surface uplift history of the terraces of the Huon Peninsula, Papua New Guinea, determined from the ^{14}C and U-series dating of wave-cut notches formed during Late Pleistocene sea level highstands. Uplift rates can be seen to have varied along the coast with much higher rates towards the south-east. (Based on J. Chappell (1974) Journal of Geophysical Research *79, Fig. 2, p. 392. Copyright by the American Geophysical Union.)*

instance, Late Pliocene – Early Pleistocene terrace deposits containing a fauna indicative of a lowland sub-tropical climate have since been elevated to a height of 4000–5000 m indicating surface uplift averaging more than 2000 m Ma^{-1}. Overall, crustal uplift rates in the Himalayas appear to be currently averaging around 5000 mm ka^{-1}. This is matched by long-term rates in some Andean ranges, such as the Cordillera Blanca in Peru which, on the basis of the exposure of a granite batholith emplaced at a depth of 8 km about 10 Ma BP, has experienced crustal uplift of around 4000–5000 m Ma^{-1} since that time.

Even these rates, though, are modest in comparison with those estimated for sections of the Southern Alps in New Zealand. This mountain range, located along the boundary of the Pacific and Indian Plates, has apparently sustained a rate of crustal uplift averaging up to 10 000 m Ma^{-1} over the past 1 Ma. This extraordinarily high uplift rate is, as in the case of the Transverse Ranges of southern California, related to the oblique convergence occurring along the associated plate boundaries (see Section 3.5.2).

For some mountain ranges the variations in local rates of surface uplift correspond well with their present-day topography; that is, the regions of highest elevation are rising at the greatest rate (Fig. 15.6). In other orogenic belts there is a mismatch between topography and present patterns of vertical movement. This is the case for the Alps where detailed relevelling surveys documenting vertical movements over the past 60 a or so show that the current maximum rates of surface uplift are to be found in parts of the alluvium filled valleys of the Rhone and the Rhine. This apparent

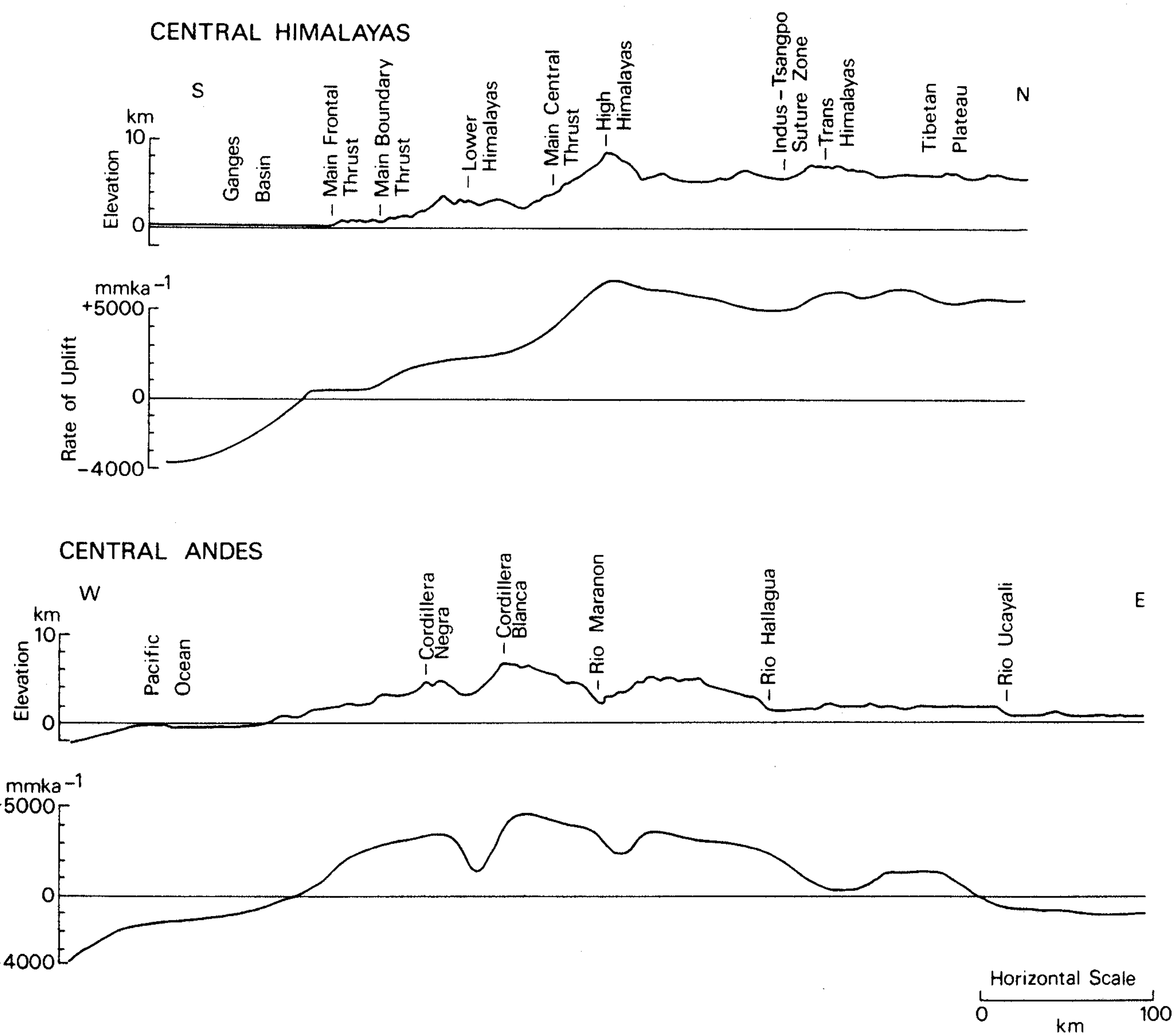

Fig. 15.6 Relationship between topography and surface uplift rates along transects across the central Himalayas and central Andes. (Modified from A. Gansser (1983) in: K. J. Hsü (ed.) Mountain Building Processes. *Academic Press, London, Fig. 2, p. 223.)*

anomaly may be at least partly explained by the continuing isostatic compensation for the removal of the ice loads occupying these valleys during the last glacial maximum.

15.2.2.2 *Epeirogenic uplift*

Perhaps because of the less obvious vertical movements involved, data on rates of epeirogenic uplift are rather sparse, at least for the long term. Rates for uplift averaged over millions of years appear to lie in the range of 10–200 m Ma^{-1}. The Colorado Plateau in the western USA, for instance, has risen about 500 m during the Late Cenozoic at a mean rate of around 100 m Ma^{-1}, while the Deccan Plateau of India has experienced uplift of 600 m or so extending over the past 40 Ma at a mean rate of about 15 m Ma^{-1}.

Although we would expect to see short-term rates that are similarly slow, this is apparently not the case. Detailed relevelling surveys carried out in the USA, for instance, demonstrate that over the past several decades there has been little to distinguish apparent rates of vertical crustal movement in the tectonically active western USA from the supposedly stable eastern part of the country (Fig. 15.7). Such data suggest remarkably high rates of surface uplift of up to 20 mm a^{-1} at the present time in regions such as Florida. These rates exceed even those in the world's most active orogenic belts so we are led to ask whether they are credible or whether they are simply an artefact of systematic errors in the levelling measurements. Although in most cases the changes in elevation recorded exceed the random error limits expected in such surveys, systematic errors certainly cannot be ruled out. On the other hand, the patterns of apparent movement observed along the eastern margin of the USA correspond to known tectonic structures and various geomorphic features such as drainage divides and stream gradient anomalies.

If such movements are real the problem remains of finding a plausible mechanism to explain them. It is obvious that such high rates in 'passive' tectonic settings of generally subdued relief cannot be sustained for long periods of time, otherwise they would generate a quite unrealistic topography (a sustained uplift of 20 mm a^{-1} would give rise to a 2000 m high mountain range in just 100 000 a assuming no erosion). This implies that such epeirogenic movements

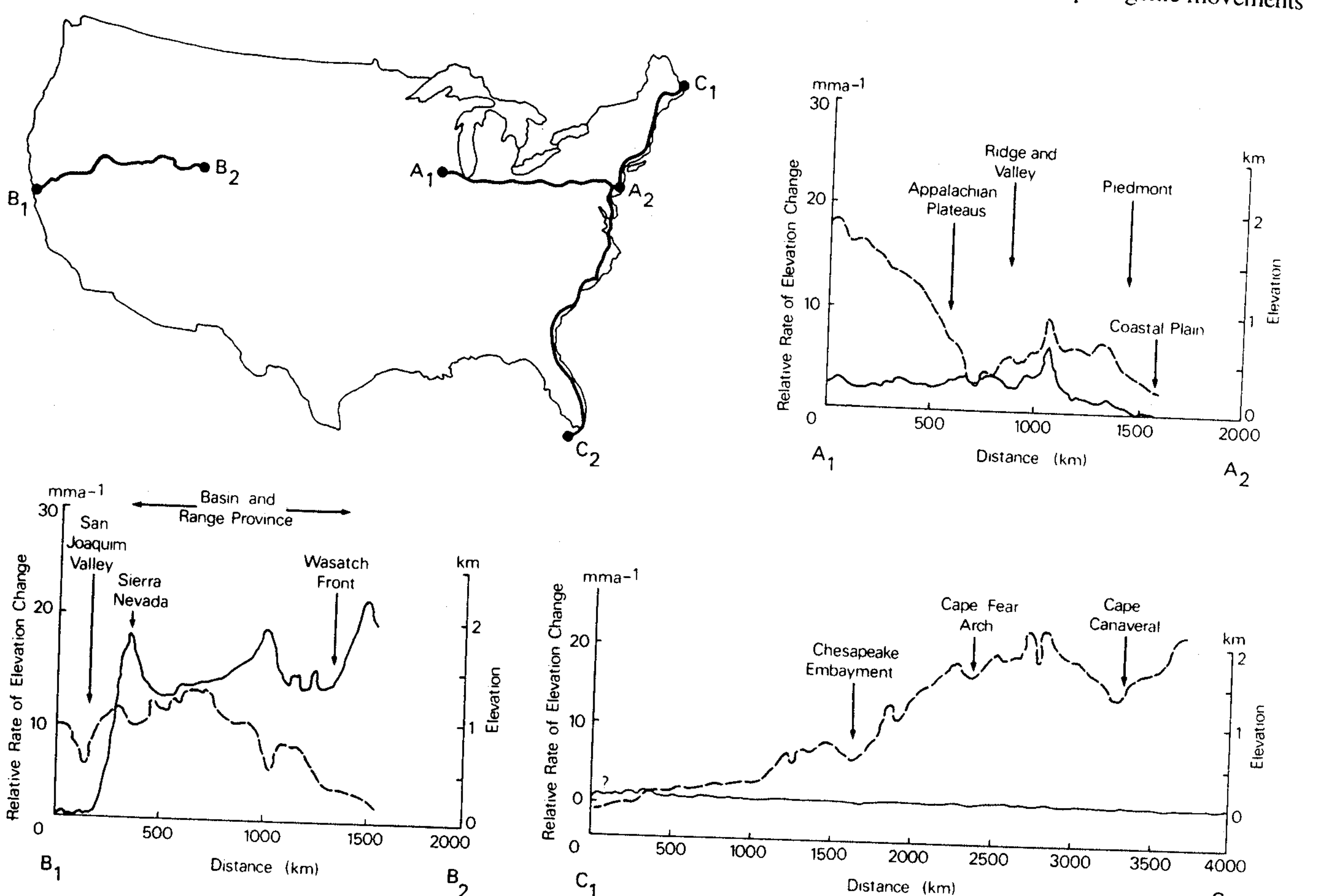

Fig. 15.7 Apparent present rates of vertical movement for various transects in the USA based on precise levelling data. Solid lines indicate topography, dashed lines apparent vertical movement. Note that rates of movement are expressed in mm a^{-1}. (Based on L. D. Brown et al. *(1980) in N.-A. Mörner (ed.)* Earth Rheology, Isostasy and Eustasy. *Wiley, Chichester, Fig. 1, p. 391 and Fig. 10, p. 399; L. D. Brown (1978)* Tectonophysics *44, Fig. 2, p. 211; and L. D. Brown and J. E. Oliver (1976)* Reviews of Geophysics and Space Physics *14, Fig. 11, p. 21.)*

are episodic or oscillatory, with periods of uplift alternating with periods of subsidence. Although episodic or oscillatory movements must be related in some way to epeirogenic mechanisms (along passive continental margins) or plate interactions (along active plate margins), the specific processes involved are unknown. Nevertheless, the considerable areal extent of the regions affected by these rapid epeirogenic movements (up to 1000 km across) implies that the cause cannot lie just in the crust but must be deep-seated and probably involves the entire thickness of the lithosphere and even possibly a part of the asthenosphere.

Although there is great uncertainty about the cause of short-term rapid uplift in many regions of the world, far more is known about the extremely high rates of uplift recorded from areas once covered by Pleistocene ice sheets. In fact the relationship between the temporal and spatial variations in rates of isostatic rebound and the thickness and location of the Pleistocene ice sheets has provided vital information about the nature of the sub-lithospheric mantle and the way in which it responds to the removal of a load. Detailed information on the decay of the great ice sheets of the last glacial and the resulting isostatic rebound of the crust is provided by raised shorelines containing datable deposits. As can be seen in Table 15.2, such evidence demonstrates that the highest rates of surface uplift are attained by crust lying below the thickest part of the ice sheet (where unloading generates the greatest degree of isostatic disequilibrium). It also shows that rates of isostatic rebound decline exponentially through time, being extremely rapid immediately after deglaciation, with rates of up to 100 000 mm ka^{-1}. Somewhat slower rates of surface uplift, at around 5000–10 000 mm ka^{-1}, continue for at least a further 15 000 a in the case of major continental ice sheets.

Table 15.2 Post-glacial rates of vertical movement (mma^{-1}) in regions covered by Late Pleistocene ice sheets

	LAURENTIDE		FENNO-SCANDIAN CENTRE	SCOTTISH CENTRE
	CENTRE	EASTERN PERIPHERY		
10–8 ka BP	100–70	70	30	10
4–3 ka BP	30	3–5	10–15	4–5
Recent decades	>(5±2) (20±5?)	−(2–3)	9–10	3.8–5.8

Note: Post-glacial eustatic rise in sea level taken into account.
Source: Data from A. A. Nikonov (1980) in: N.–A. Mörner (ed.) *Earth Rheology, Isostasy and Eustasy*. Wiley, Chichester, Table 2, p. 347 compiled from various sources.

15.3 Present fluvial denudation rates

The analysis of contemporary rates of denudation raises major difficulties, but it provides the only means we have of determining which factors are most significant in controlling the rate at which landscapes are eroded. An assessment of fluvial denudation rates is important because the output of sediment and dissolved constituents from the usually well-defined boundaries of a drainage basin enables us to monitor the net effects of weathering, slope processes and fluvial transport across a specific area. Defining source regions for the removal of material by glaciers, and especially wind, is far more difficult. Perhaps the greatest limitation in using present-day denudation rates as a basis for understanding long-term landform development is that we cannot be sure how typical they are. This is why it is also vital to examine other evidence for long-term rates of denudation.

15.3.1 Methods of Measurement and Estimation

The basis for determining fluvial denudation rates is the estimation of the **solid** (sediment) and solute (dissolved) load carried by rivers (see Section 8.4.1). A very approximate estimate can be achieved simply by multiplying the mean sediment and solute concentration calculated from a small number of samples by mean discharge. More accurate estimates, however, must take into account the way in which the concentrations of sediment and dissolved constituents vary with discharge and, in particular, the way in which they vary with flood events. This is accomplished by using **sediment** and **solute rating curves** constructed from equations which describe the best fit relationships between sediment and solute concentrations and discharge. For greater accuracy separate rating curves can be used for rising or falling stage relationships and for seasonal flows. Solid and solute transport rates can then be calculated by relating the rating curve to either continuous stream-flow data, or flow-duration curves based on hourly, daily or even monthly data. Such a procedure is subject to errors, particularly where the frequency of sampling is low. In small basins transport rates can vary significantly from hour to hour, and even in large basins a high degree of variability can occur between individual years.

Since data on sediment and solute transport rates are not available for all drainage basins, estimates of denudation rates on a continent-wide or global basis must be founded on some form of extrapolation. This is usually based on empirical relationships observed between measured solid and solute transport rates and the factors thought to control these rates, especially those related to climate and relief. Where such relationships are found to be strong it is then possible to estimate the sediment and solute transport rates on the basis of climate and relief, and, in some cases, other variables.

This approach has been much favoured by some French geomorphologists. For instance, in an extensive survey published in 1960, F. Fournier estimated the global pattern of denudation by relating sediment yield data on 78 basins (ranging in area from 2460 to 1 060 000 km^2) in a variety of climatic zones to a seasonality of rainfall index. This was expressed as p^2/P, where p is the mean monthly maximum precipitation and P is the mean annual precipitation. The idea of such a seasonality of rainfall index is that it indicates the importance in sediment transport of peak discharges generated by intense seasonal rainfall. Fournier also incorporated the effect of relief by using separate regression equations for different relief types. Because the relationships between sediment and solute transport rates and the various climatic and relief variables used are not perfect, such an approach can only give an approximate picture of the variation in denudation rates.

Additional information on solid and solute loads for major river basins is constantly accumulating, and most estimates of contemporary global denudation are now increasingly based on these larger data sets rather than extrapolation from results for a limited number of basins. Nevertheless, there remain significant difficulties in converting solid and solute load data into meaningful estimates of denudation rates. A general problem is simply the unreliability of measurements for many of the world's major rivers. The data on solid and solute load are inadequate for basins such as the Amazon, Ganges and Brahmaputra which contribute large quantities of both sediment and dissolved constituents to the world's oceans. The data for Chinese rivers, which are also important sediment contributors, also used to be poor but are now much improved following the establishment of comprehensive monitoring programmes for several major basins. In some cases it is even difficult to find out details of the method of measurement used. A widely cited estimate of the sediment yield for the Irrawady River in Burma is in fact based on a few measurements taken by unknown methods in the 1870s. Even where modern techniques of measuring discharge and of sampling sediment and solute concentrations have been applied, there may still be significant errors in rating curves and insufficient information on the monthly and annual variability of discharge and the occurrence of rare peak discharges which may be particularly important in affecting the overall rate of sediment transport. Moreover, if there is no gauging station at the mouth of a river the total load transported to the ocean may be overestimated as there may be significant deposition or precipitation downstream of the sampling point. Other important problems relate specifically to the accurate estimation of solid and solute load, and we will now examine these in some detail.

15.3.1.1 *Solid load*

In most measurements of solid load transport only the suspended sediment is sampled and, because of the difficulties involved with large rivers, no attempt is made to measure bed load. In lowland tropical rivers this may not be a major problem; for example it has been estimated that bed load accounts for only 1 per cent of the total sediment load of the Amazon River. In mountainous terrains, though, bed-load transport can be very significant and in such environments the widely applied assumption that bed load is about 10 per cent of suspended load may be a drastic underestimate.

Table 15.3 Comparison of sediment yields (t km^{-2} a^{-1}) under natural and artificial conditions in various countries

LOCATION	NATURAL	CULTIVATED LAND	BARE SOIL
UK	10–50	10–300	1000–4500
USA	3–300	500–17 000	400–9000
China	<200	15 000–20 000	28 000–36 000
India	50–100	30–2000	1000–2000
Nigeria	50–100	10–3500	300–15 000
Ivory Coast	3–20	10–9000	1000–75 000

Source: Data from R. P. C. Morgan (1986) *Soil Erosion and Conservation*. Longman, London, Table 1.1, p. 5, based on various sources.

Another problem in estimating natural rates of denudation and in evaluating the factors that control them is that in many river basins sediment yields have been changed dramatically, at least at the local scale, through anthropogenic effects. Dam construction promotes a downstream reduction in sediment yield, but cultivation, surface mining and urban construction all lead to disturbance of the soil and consequently can give rise to sediment yields far in excess of those generated from naturally vegetated surfaces (Table 15.3). It has been estimated, for instance, that the conversion of forest to cropland in the states of the mid-Atlantic seaboard of the USA has led to a tenfold increase in sediment yield, but such increases within a basin are not necessarily reflected immediately by an increase in sediment load at the basin terminus. In the case just cited, for example, it has been calculated that over 90 per cent of the sediment stripped from the upland regions of the Piedmont region of the south-east USA remains stored as colluvium on hillslopes, and as alluvium in valley bottoms. In other words there is not a steady state between sediment supply and sediment transport. A similar situation exists in the upper Mississippi Basin, where a detailed sediment budget study has demonstrated that only 7 per cent of the sediment being removed from the upper parts of a tributary valley is currently reaching the Mississippi River (Fig. 15.8).

The figure of 7 per cent indicates the proportion of sediment being mobilized within the basin that actually reaches the basin outlet and represents the **sediment delivery ratio** for the basin. This is a central concept in understanding the dynamics of drainage basin erosion through time. In the long term the sediment delivery ratio must be approximately unity otherwise the river basin would become progressively choked with sediment. (In very large basins it may be less than unity, since sediment may progressively accumulate in the basin as the basin floor subsides under the load of overlying sediment.) In the short term, however, there

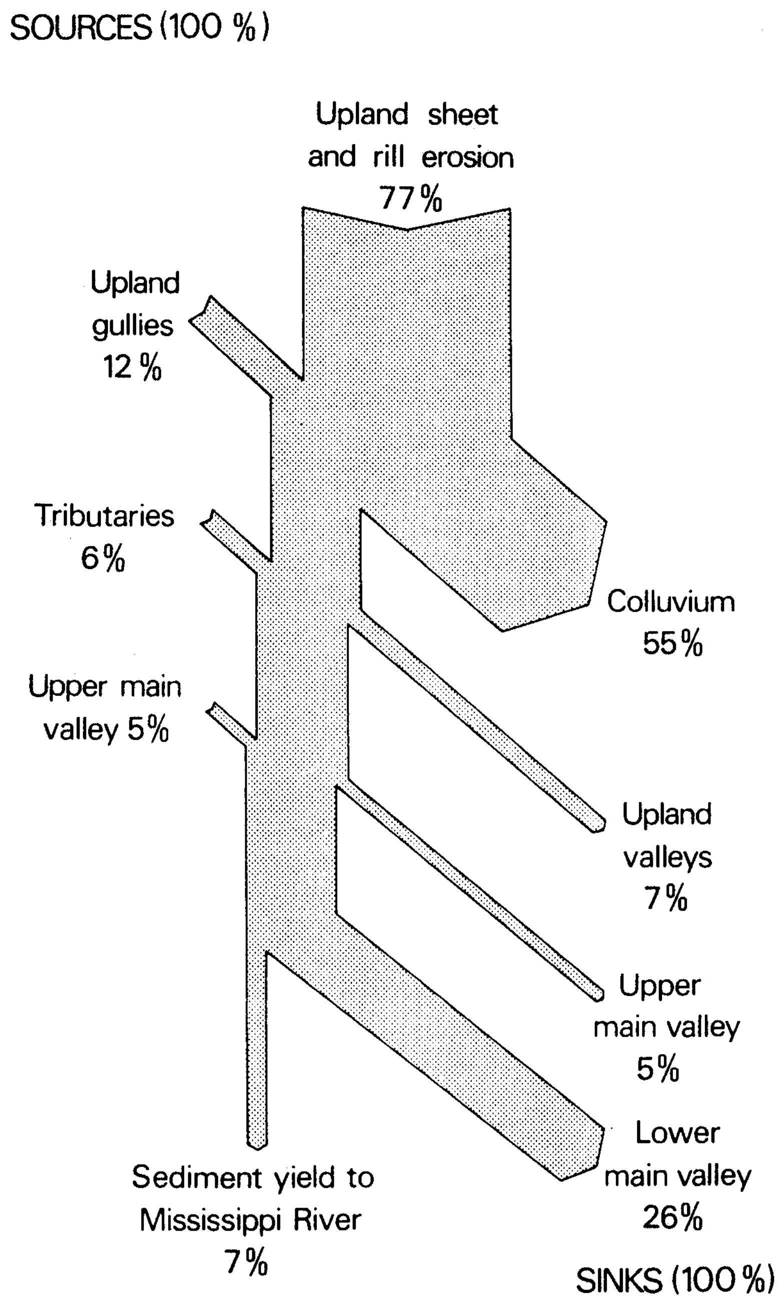

Fig. 15.8 Sediment budget from 1938–75 for Coon Creek, a tributary of the Mississippi River in Wisconsin, USA. The figures show the percentage of sediment mobilized from different sources within the basin and the percentage deposited in different temporary storages (sinks). (Based on data from S. W. Trimble (1983) American Journal of Science *283, 454–74.)*

may be transfers of material within a drainage basin but little export of material from it. Debris removed from slopes can be stored in fans, talus slopes and as river terraces and floodplain sediment.

The disparity between the amount of erosion occurring on hillslopes within a basin and the total amount of sediment being removed from it will tend to become more marked as drainage basin area increases, since larger catchments are likely to have a greater channel storage capacity in the form of extensive floodplains. An additional factor will be basin morphology with mountainous catchments generally having a smaller sediment storage capacity than lowland basins. Eventually this stored sediment will be mobilized and transported out of the basin, but this may take decades or centuries, or, in the case of very large basins, many thousands of years or more. Glacial and talus deposits generated in mountain valleys during the Late Pleistocene have still not been removed from many mid-latitude basins, while in the Amazon Basin only the major rivers are currently able to rework the coarser alluvium that was deposited during the more arid phases of the Quaternary.

15.3.1.2 Solute load

In view of the complexity of measuring separately each dissolved constituent in stream water, the easily measured electrical conductivity, or **specific conductance**, of the water is often used to provide an estimate of the solute concentration. Although there is a strong correlation between the concentration of ionic species in solution and electrical conductivity, the exact relationship varies depending on the concentrations present of particular dissolved constituents. Moreover, SiO_2, which is a significant component of many tropical lowland rivers, is not recorded by this technique. The solute load of rivers can also be enhanced by anthropogenic inputs. This is especially the case in basins containing industrial activities, and also in agricultural regions where the fertilizers applied to crops can find their way into stream waters in significant quantities. Human activities can also lead to an increase in rates of chemical weathering and thereby enhanced rates of solute input into stream waters.

The most important problem in estimating the contribution of solute transport to total denudation is the separation of the denudational and non-denudational components. Allowances must be made for dissolved constituents introduced into a basin through precipitation, since these represent a non-denudational component additional to the solutes released through bedrock weathering. An adjustment can either be made by using a knowledge of the chemical and mineralogical composition of the lithologies exposed in a basin to predict the solutes likely to be derived from bedrock weathering or, for greater accuracy, the chemical composition of the precipitation can be measured directly.

Table 15.4 gives the average composition of world river water and shows that two major non-denudational components introduced by precipitation are Na^+ and Cl^-, the two most abundant dissolved constituents in sea water. An even more important non-denudational component, however, is HCO_3^- which arises from the incorporation of atmospheric CO_2 during weathering reactions. Together these non-denudational components on average amount to about 40 per cent of the solute load of rivers but there are major departures from this figure. In regions of rapid rock weathering in highly reactive lithologies the non-denudational component is relatively minor, but in some tropical lowland rivers draining thoroughly leached and almost chemically inert weathering mantles the non-denudational component may be very significant. In the Rio Ucayali in the mountainous Andean region of the Amazon Basin it is estimated that only 4.8 per cent of Na^+, K^+, Ca^{2+} and Mg^{2+} comes from precipitation but in the Rio Tefé, a lowland tributary of the Amazon, 81 per cent of these four constituents is contributed by precipitation. Apart from their SiO_2 content the chemical composition of many of the lowland Amazon tributaries is in fact very similar to extremely dilute sea water, a reflection of the predominance of precipitation-derived solutes. An additional consideration relevant to many different environments and time scales is that the denudational component of the solute load can also be underestimated if solutes taken up in vegetation are removed from the basin as litter in streams.

15.3.1.3 Estimation of volumetric changes

A rather neglected problem concerning the estimation of denudation rates from solid and solute load data is the conversion of a measurement expressed as a mass per unit area per unit time (usually t km^{-2} a^{-1}) to a volumetric equivalent. Since we are concerned with the change in the *form* of the ground surface through time, it is the volumetric change which is of importance. At first sight this seems a trivial task: we simply divide the mass of material by the density of the bedrock (say 2700 kg m^{-3} for rocks making up most of the upper continental crust) to give a volume (m^3 km^{-2} a^{-1}). We then convert this to an *average* rate of lowering for the whole basin, acknowledging, of course, that the actual change in elevation will inevitably vary considerably from place to place. Conveniently, a volume change express-

Table 15.4 Average composition of world river water and estimates of denudational and non-denudational contributions for different constituents

	Ca^{2+}	Mg^{2+}	Na^+	K^+	Cl^-	SO_4^{2-}	HCO_3^-	SiO_2	Total
Average composition of world river water (concentration (mg l^{-1}))	13.5	3.6	7.4	1.35	9.6	8.7	52.0	10.4	106.6
Provenance of major solute components (%)									
Non-denudational:									
Precipitation (oceanic salts)	2.5	15	53	14	72	19	—	—	12
Atmospheric CO_2	—	—	—		—	—	57	—	28
Denudational:									
Chemical weathering	97.5	85	47	86	28	81	43	100	60

Source: Based largely on data in M. Meybeck (1983) in: *Dissolved Loads of Rivers and Surface Water Quantity/Quality Relationships* International Association of Hydrological Sciences Publication **141**, 173–192.

ed in $m^3 km^{-2} a^{-1}$ is equivalent to a mean rate of ground lowering expressed as mm ka^{-1}. The averaging over a period of 1000 a is largely conventional and has the advantage of giving an amount of lowering that can be readily appreciated. It does not imply that the rate of denudation has been measured over this length of time, and this is, of course, not the case when using stream load data.

Unfortunately, this simple procedure does not necessarily yield an accurate estimate of the equivalent average volume change actually occurring in the landscape. If soil or weathered material, which may have a bulk density between 1100 and 2000 kg m^{-3}, is being eroded, is it appropriate to use the density of the underlying bedrock in calculating the associated volume change? The answer is yes if the rate of rock weathering and removal of weathered material and soil are more or less in a steady state, but our understanding of the mode of landscape modification in many different geomorphic environments, and especially those of lowland tropical regions of subdued relief, suggests that periods characterized by low denudation rates, during which the weathering mantle increases in depth, are punctuated by phases of active erosion. In such cases rock weathering and weathering mantle erosion are not in equilibrium, and in this situation it is not clear that converting sediment removal into a rate of ground lowering on the basis of the original bedrock density is meaningful.

This problem is compounded when we consider chemical denudation. Conventionally, the specific solute load of streams measured in t $km^{-2} a^{-1}$ is converted into a rate of ground lowering in mm ka^{-1} using an appropriate bedrock density.

Table 15.5 Estimates of total transport by rivers of solids and solutes to the oceans and equivalent estimated denudation rates

AUTHOR	MEAN LOAD (10^9 t a^{-1})	MEAN LOAD (t $km^{-2} a^{-1}$)	EQUIVALENT DENUDATION RATE† (mm ka^{-1})
*Solid load**			*Mechanical*
Fournier (1960)	58.1	392.6	145.4
Jansen and Painter (1974)	26.7	180.4	66.8
Schumm (1963)	20.5	138.5	51.3
Holeman (1968)	18.3	123.6	45.8
Milliman and Meade (1983)	13.5	91.2	33.8
Lopatin (1952)	12.7	85.8	31.8
Solute load			*Chemical‡*
Goldberg (1976)	3.9	26.4	5.9
Livingstone (1963)	3.8	25.7	5.7
Meybeck (1979)	3.7	25.0	5.6
Meybeck (1976)	3.3	22.3	5.0
Alekin and Brazhnikova (1960)	3.2	21.6	4.8

* Suspended load only.
† Denudation rates based on a rock density of 2700 kg m^{-3}.
‡ Rates for chemical denudation assume that 40% of total solute load is from non-denudational sources.

Where solutes come from the direct solution of bedrock, as occurs to a large extent in limestone terrains, this is a valid procedure. But in many cases bedrock weathering takes place without any change in volume. Where such isovolumetric weathering occurs (see Section 6.2.4.1) solutes are lost to stream waters, but there are no associated volume changes as the transformation of fresh bedrock into saprolite composed largely of clay minerals is accompanied by a compensating decrease in bulk density. Even where weathering is not strictly isovolumetric there is almost invariably a significant compensation through a decrease in bulk density associated with the removal of dissolved constituents from the weathering mantle. Consequently, in lowland regions mantled by thick weathering profiles, it is misleading to view the removal of weathering-derived solutes *in the short term* as necessarily being directly related to chemical denudation in the sense of a reduction of the average elevation of a basin. In considering the rates of mechanical and chemical denudation discussed in the following section, it will be important to keep these points in mind.

15.3.2 Rates of mechanical and chemical denudation

Table 15.5 lists a range of estimates of total solid and solute load transport to the world's oceans and gives the equivalent global mean denudation rates (assuming an average source rock density of 2700 kg m^{-3}). It is important to point out here that these, and all other estimates of denudation rates, are subject to variable, and often large, errors. Although rates may be reported to a precision of one decimal place, this certainly does not imply that they are accurate to that degree.

The estimates in Table 15.5 are based on the total land area of 148×10^6 km^2 and would be increased by about 40 per cent if only the global area of external drainage (about 105×10^6 km^2) is considered. The various estimates in Table 15.5 are not directly comparable as the basis for their calculation differs, but it appears that the early estimate of Fournier, based largely on his suggested relationship between sediment yield and seasonality of precipitation, is too high. The more recent low estimate of Milliman and Meade, on the other hand, is for actual solid load transport to the oceans at the present day, and includes the effects of sediment entrapment by dams on rivers such as the Colorado, Nile and Zambezi.

What, then, is the best estimate we can make of current global denudation rates excluding the effects of human activities? Taking the Milliman and Meade estimate of 13 500 Mt a^{-1} for suspended sediment transport, we can add on 500 Mt a^{-1} as a reasonable estimate of the material trapped by dams and 1500 Mt a^{-1} to allow for unrecorded bed-load transport. In addition we perhaps should allow 500 Mt a^{-1} as a deduction to take account of the increase in erosion rates as a consequence of human activities. These are admittedly rough approximations, but the total this gives of 15 000 Mt a^{-1} is probably a reasonable estimate of natural solid load transport to the oceans.

Estimates for global solute transport to the oceans are less variable, and we will take the recent figure established from a detailed global survey by M. Meybeck of 3700 Mt a^{-1} as being the most accurate available. From this we need to make a deduction to allow for the non-denudational component. As a global average 40 per cent is probably a reasonable figure and this gives an estimate of 2200 Mt a^{-1} for 'denudational' solute load transport. This gives a global mean annual transport of 17 200 Mt of material to the oceans which, averaged over the entire land area of the continents, gives a rate of 116 t km^{-2} a^{-1}. Assuming a mean source rock density of 2700 kg m^{-3}, this converts to a mean global denudation rate of 43 mm ka^{-1}. If we exclude areas of internal drainage where material is being transported but not removed from the continents, this figure rises to 61 mm ka^{-1}. Of this total about 85 per cent is accounted for by solid load transport and 15 per cent by the transport of solute load. It is reasonable to assume that, over large areas, denudation will lead to a compensatory isostatic rebound which reduces the 43 mm ka^{-1} mean rate of lowering with respect to sea level to only 8 mm ka^{-1}.

Such a global mean conceals great variations from area to area and basin to basin. Around 70 per cent of the total load transported to the oceans is provided from only 10 per cent of the land area and just three rivers, the Ganges, the Brahmaputra and the Huang He (Yellow) carry 20 per cent of the global fluvial sediment load. Figure 15.9 shows the results of a global survey of sediment yields based on measurements from more than 1500 sites and indicates that specific sediment yield rates exceed 1000 t km^{-2} a^{-1} (roughly equivalent to a mechnical denudation rate of 370 mm ka^{-1}) in the world's mountainous regions. By contrast, rates are less than 50 t km^{-2} a^{-1} (equivalent to approximately 19 mm ka^{-1}) in many lowland regions.

If we examine the world's largest drainage basins we find that present-day total denudation rates range from 3 mm ka^{-1} for the interior Chari Basin in Africa and 5 mm ka^{-1} for the Kolyma Basin in eastern Siberia, up to 529 mm ka^{-1} for the Huang He Basin of China and 677mm ka^{-1} for the Brahmaputra Basin draining the eastern Himalayas (Table 15.6, Fig. 15.10). These figures, however, by no means represent the extremes found in smaller basins. Minimum sediment yields lie well below 2 t km^{-2} a^{-1} (equivalent to <1 mm ka^{-1}) (1.7 t km^{-2} a^{-1}, for instance, for the Queanbeyan Basin (172 km^2) in the southern tablelands of the East Australian Highlands and less than 1 t km^{-2} a^{-1}

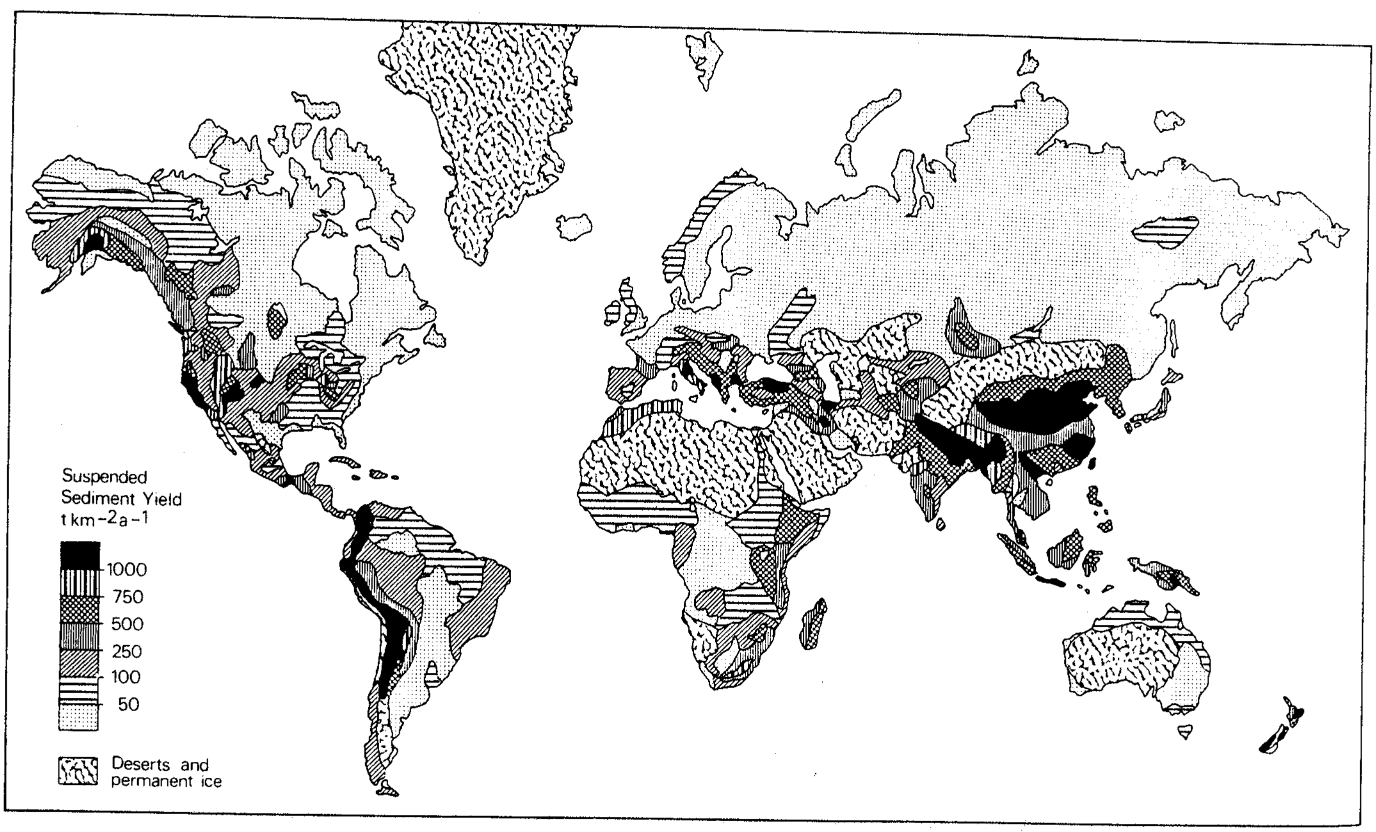

Fig. 15.9 Global pattern of yields of suspended sediment. The values relate to intermediate-sized basins of 10^4–10^5 km^2. (After D. E. Walling and B. W. Webb (1983) in: K. J. Gregory (ed.) Background to Palaeohydrology. *Wiley, Chichester, Fig. 4.2, p. 76.)*

Table 15.6 Estimated denudation rates for the world's thirty-five largest drainage basins based on solid and solute transport rates

	DRAINAGE AREA (10^6 km^2)	RUNOFF (mm a^{-1})	TOTAL DENUDATION (mm ka^{-1})	MECHANICAL DENUDATION (mm ka^{-1})	CHEMICAL* DENUDATION (mm ka^{-1})	CHEMICAL DENUDATION AS % OF TOTAL
Amazon	6.15	1024	70	57	13	18
Zaire (Congo)	3.82	324	7	4	3	42
Mississippi	3.27	177	44	35	9	20
Nile	2.96	30	15	13	2	10
Paraná (La Plata)	2.83	166	19	14	5	28
Yenisei	2.58	217	9	2	7	80
Ob	2.50	154	7	2	5	70
Lena	2.43	206	11	2	9	81
Chiang Jiang (Yangtze)	1.94	464	133	96	37	28
Amur	1.85	175	13	10	3	22
Mackenzie	1.81	169	30	20	10	33
Volga	1.35	196	20	7	13	64
Niger	1.21	159	24	13	11	47
Zambezi	1.20	186	31	28	3	11
Nelson	1.15	96	—	—	—	—
Murray	1.06	21	13	11	2	18
St Lawrence	1.03	434	13	1	12	89
Orange	1.02	89	58	55	3	5
Orinoco	0.99	1111	91	78	13	14
Ganges	0.98	373	271	249	22	8
Indus	0.97	245	124	108	16	13
Tocantins	0.90	385	—	—	—	—
Chari	0.88	69	3	2	1	29
Yukon	0.84	232	37	27	10	28
Danube	0.81	254	47	31	16	35
Mekong	0.79	595	95	75	20	21
Huang He (Yellow)	0.77	63	529	518	11	2
Shatt-el-Arab	0.75	61	104	93	11	11
Rio Grande	0.67	5	9	6	3	38
Columbia	0.67	375	29	16	13	46
Kolyma	0.64	111	5	3	2	31
Colorado	0.64	31	84	78	6	7
São Francisco	0.60	151	—	—	—	—
Brahmaputra	0.58	1049	677	643	34	5
Dnepr	0.50	104	6	1	5	88

* Allowance made for non-denudational component of solute loads.
Source: Based primarily on data from M. Meybeck (1976) *Hydrological Sciences Bulletin* **21**, 265–89, and J. D. Milliman and R. H. Meade (1983) *Journal of Geology* **91**, 1–21.

for several rivers in Poland). Maximum sediment yields exceed 10 000 t km^{-2} a^{-1}, with the Haast River draining a region of high precipitation and very rugged relief in the Southern Alps in New Zealand having a yield of 12736 t km^{-2} a^{-1} (equivalent to 4717 mm ka^{-1}). But even this is exceeded by the 53 500 t km^{-2} a^{-1} of the Huangfu-chuan River, a tributary of the Huang He, which drains over 3000 km^2 of gullied loess-covered terrain in a region of sparse vegetation with a semi-arid climate characterized by periodic intense storms. This is equivalent to a denudation rate of 19 814 mm ka^{-1}, a figure which will only be sustained while supplies of readily erodible loess remain in the catchment.

Chemical denudation rates are in general less variable than those for mechanical denudation, but none the less still exhibit a wide range. Minimum solute load yields lie below 1 t km^{-2} a^{-1}, whereas maximum yields of 6000 t km^{-2} a^{-1} occur in rare instances where rivers drain highly soluble deposits such as halite. More usual maxima lying below 1000 t km^{-2} a^{-1} (equivalent to 370 mm ka^{-1}) occur in limestone regions. High rates of chemical denudation are invariably observed in humid mountainous regions. In such environments rates of chemical weathering are high as the weathering front is at, or close to, the surface, but the accompanying high rates of mechanical denudation prevent the accumulation of thick weathering profiles (see Section 6.2.4.1). Conversely, minimum rates are recorded in semi-arid regions where runoff is very low (although solute concentrations may be very high), in lowland humid tropical regions where solute *concentrations* are generally extremely low and in high latitude lowland terrains where both runoff and solute concentrations are low.

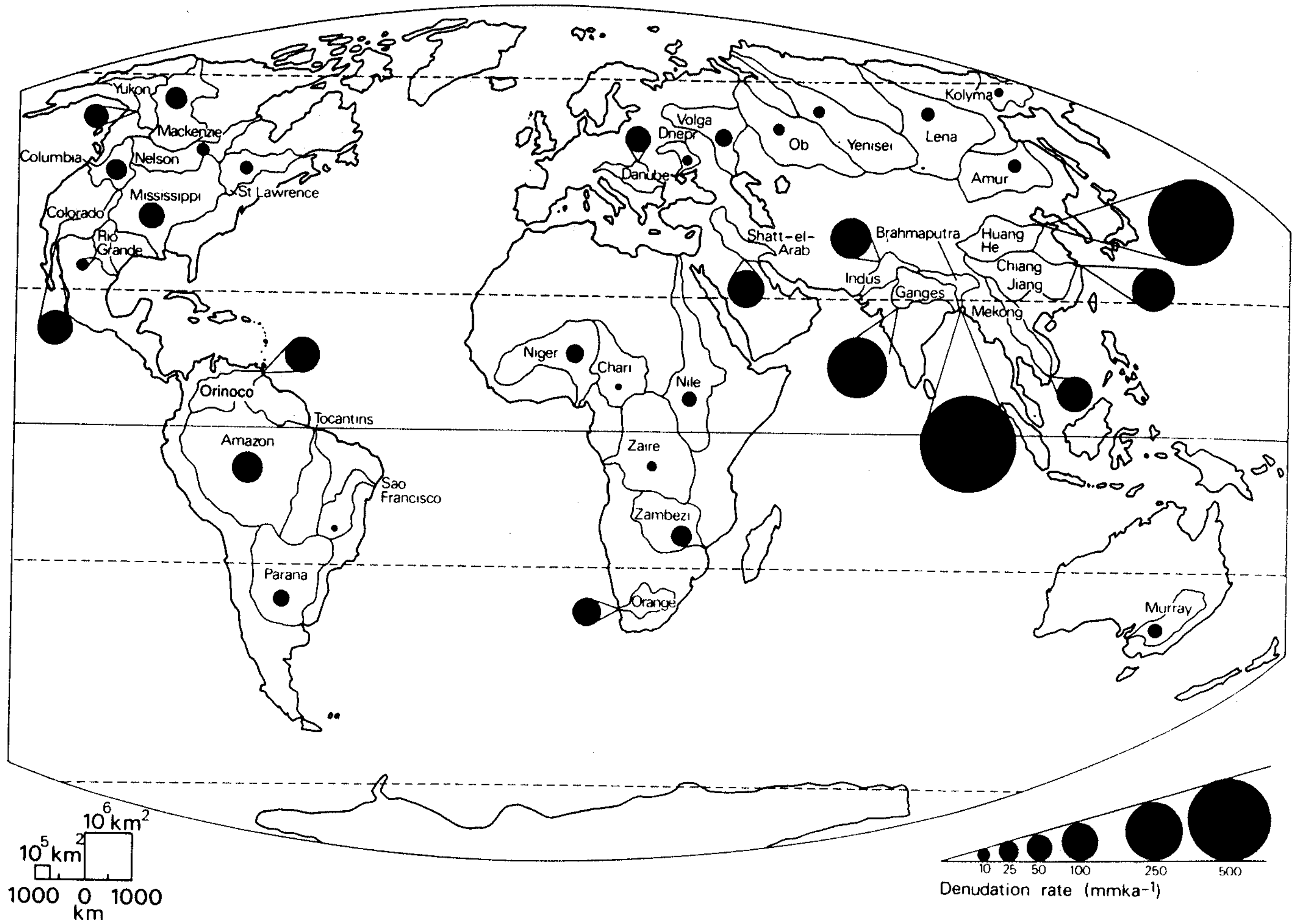

Fig. 15.10 *Denudation rates for the world's 35 largest drainage basins based on solid and solute load data. Allowance has been made for the non-denudational component of solute loads. Source rock density is assumed to be 2700 kg m^{-3} (Based on data in Table 15.6.)*

15.3.3 Relative importance of mechanical and chemical denudation

On the global scale we have already noted that the ratio of mechanical to chemical denudation is about 6 : 1, but what are the variations about this world mean? It is evident from Table 15.6 that for some basins, especially those in a predominantly humid lowland environment such as the great Siberian basins of the Ob, the Yenisei and the Lena, chemical denudation can actually greatly exceed mechanical denudation (although we must keep in mind the point made in Section 15.3.1.3). The other extreme is reached with those basins with extremely high sediment yields such as the Brahmaputra and the Huang He where chemical denudation represents 5 per cent or less of total denudation. Overall chemical denudation rates are less variable than those for mechanical denudation, ranging over two orders of magnitude rather than three or more. Rates of mechanical and chemical denudation overall show a fairly strong positive relationship, but as the ratio of solid to solute load per unit area increases as total denudation increases, chemical denudation becomes *proportionally* less significant in drainage basins experiencing higher total denudation rates. This is apparent in Fig. 15.11 which illustrates the relative and absolute rates of solid and solute load transport for major basins. It is apparent that those rivers transporting the greatest solid load are in general those also transporting the highest solute load. Note also that the solute load tends to form a greater proportion of total load when the total load is small.

A fairly clear pattern emerges from Figure 15.11 which suggests the primary influences on relative rates of mechanical and chemical denudation. The highest rates for both are found in basins draining major orogenic belts such as the Brahmaputra and the Ganges. The Chiang Jiang (Yangtze) Basin, on the other hand, has a particularly high rate of chemical denudation presumably because it occupies large areas of limestone terrain in Szechwan Province in western China. Rivers draining largely semi-arid regions such as the Colorado, Orange and Shatt-el-Arab (Tigris and Euphrates)

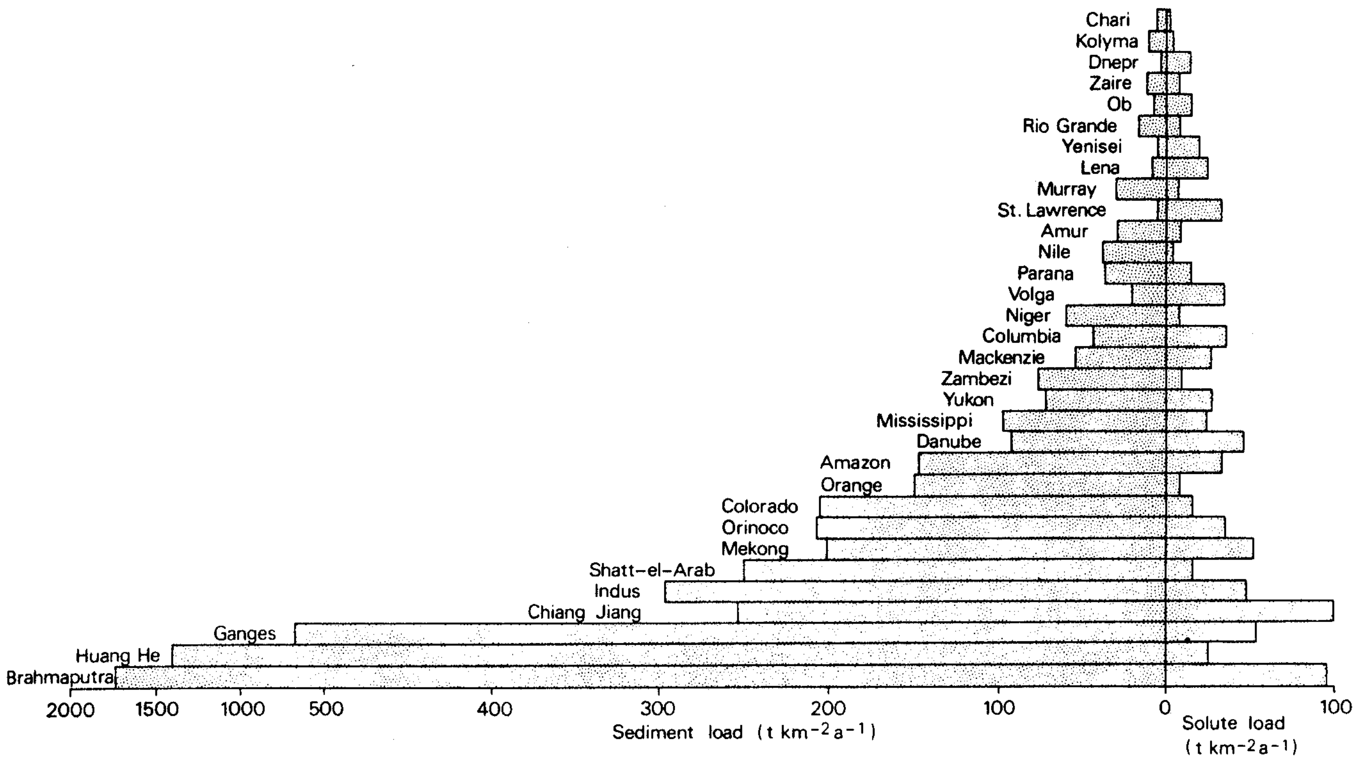

Fig. 15.11 Sediment and solute loads for the world's largest drainage basins. Solute loads represent the estimated denudational component only. Data for the Nelson, Tocantins and São Francisco Basins are not available. (Based primarily on data in M. Meybeck (1976) Hydrological Sciences Bulletin *21, 265–89 and J. D. Milliman and R. H. Meade (1983)* Journal of Geology *91, 1–21.)*

can be seen to have very low rates of solute transport in relation to total transport. This contrasts starkly with the basins of humid and subarctic regions such as the Lena, Yenisei, Ob and Dnepr which have high rates of solute transport in relation to total transport. The extremely high relative rate of solute transport for the St Lawrence Basin (some 87 per cent of total transport) is explained, at least in part, by the large proportion of sediment currently being trapped in the Great Lakes.

It is important to note that even with its predominantly semi-arid climate the Colorado Basin still has a higher rate of chemical denudation than either the humid tropical Zaire Basin or the largely subarctic Ob Basin. This is presumably because of the high rates of chemical denudation occurring in the high relief section of the Colorado Basin in the Rocky Mountains and exemplifies the generalization that, as for mechanical denudation, rates of chemical denudation appear to be much more strongly related to relief than to climate. This point is graphically illustrated in Figure 15.12 which compares the solid and solute loads of major tributaries of the Amazon. The Marañón and Ucayali Basins in the mountainous Andean part of the Amazon Basin show high rates of both solid and solute transport in comparison with the Negro Basin which drains the plateau country of the Guiana Highland, and the Xingu Basin which occupies a region of very low relief on the northern margin of the Mato Grosso. Some 85 per cent of the total solute load of the Amazon Basin originates from the Andean region of the catchment with the Marañón and Ucayali Basins alone jointly contributing 45 per cent.

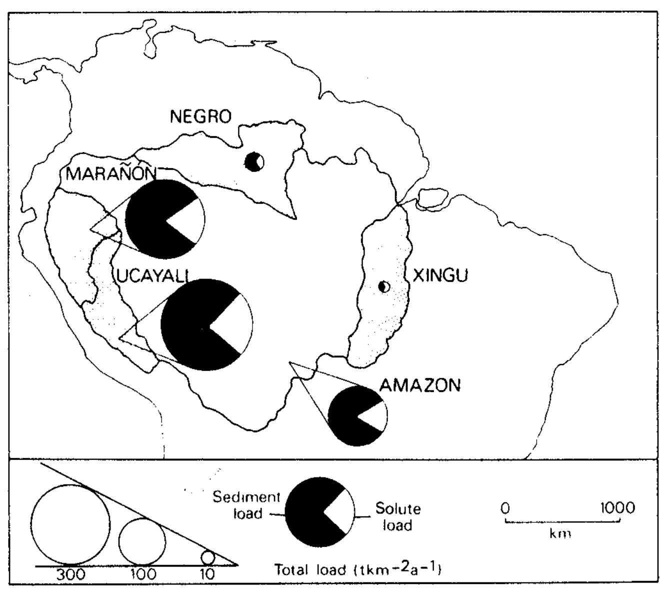

Fig. 15.12 Rates of solid and solute load transport in sub-catchments of the Amazon Basin. Rates of solute load transport are for the estimated denudational component (Based on data in M. Meybeck (1976) Hydrological Sciences Bulletin *21, 265–9.)*

Although we expect that there will frequently not be a

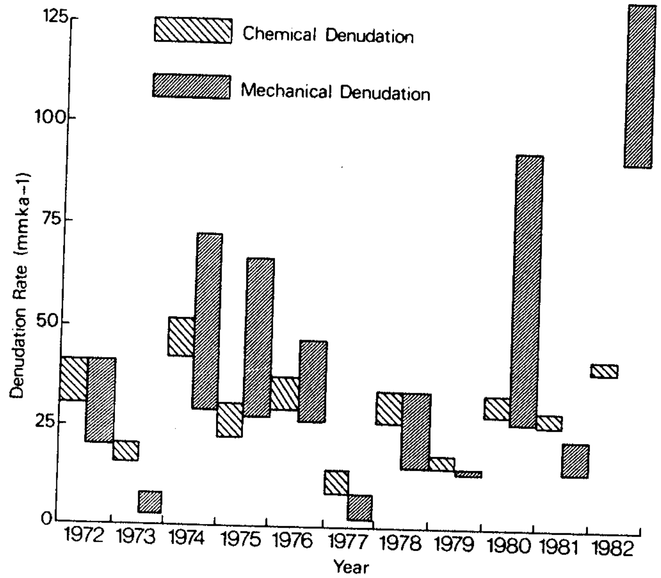

Fig. 15.13 Annual range and variability in rates of mechanical and chemical denudation for four catchments in Idaho, USA. (Based on J. L. Clayton and W. F. Megahan (1986) Earth Surface Processes and Landforms *11, Fig. 3, p.394.)*

steady state between sediment supply and sediment transport rates, in general this is unlikely to be the case for the release of solutes and their transport out of the basin. This point needs to be emphasized since it must be taken into account in any comparison of rates of mechanical and chemical denudation. Another factor which has to be considered, especially when comparisons are based on short-term records, is the different ways in which sediment and solute transport rates vary through time. This is well illustrated by a detailed study of four small forested catchments on a coarse-grained granitic lithology in Idaho, USA (Fig. 15.13). Here the mean total denudation rate over an eleven year period was calculated to be 8.9 mm ka^{-1}. The 11 a mean mechanical denudation rate was found to exceed that for chemical denudation for three out of the four catchments. Nevertheless, because of the greater temporal variability of mechanical denudation it was more probable that in any one year chemical denudation would exceed mechanical denudation in three out of the four basins. In this case episodic high rates of mechanical denudation were considered to be related to high peak flows generated by snow-melt runoff after winters with heavy snowfalls.

15.4 Long-term fluvial denudation rates

15.4.1 Methods of estimation

The major problem with using sediment and solute load data as a basis for estimating long-term denudation rates is that we cannot be certain that we are not sampling an atypical period and therefore extrapolating from an unrepresentative sample. In particular, we have already highlighted the problems of anthropogenic influences on present-day sediment and solute load data, as well as the complications arising from the lag between upstream sediment supply and sediment removal from basins. Fortunately, there are a number of other indirect means which we can use to estimate long-term rates of denudation.

15.4.1.1 Estimates from sediment volumes

The variability of sediment yields over periods of a few decades even in cool, humid temperate environments is evident from studies of sedimentation rates in small lakes and reservoirs. Denudation rates can be estimated on the basis of a known volume of sediment deposited over a known period of time and originating from a known source

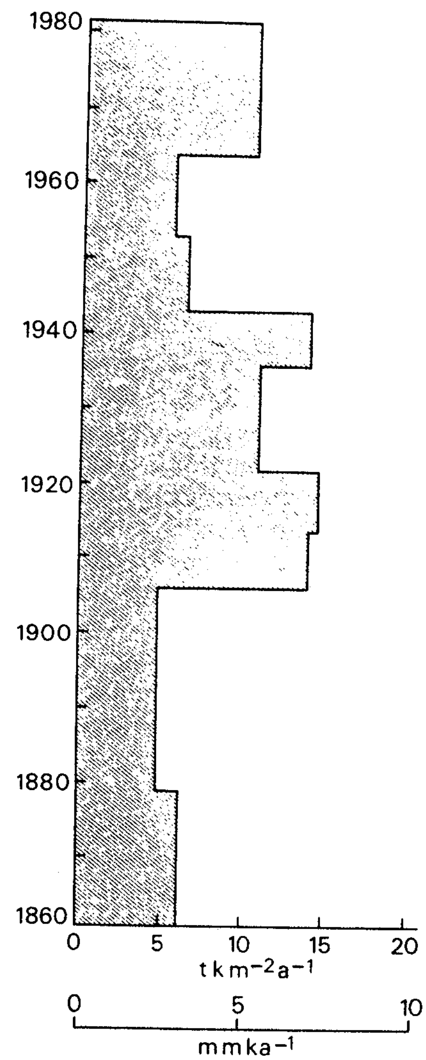

Fig. 15.14 *Temporal variations in the rate of mechanical denudation in the Merevale catchment, Warwickshire, UK, based on rates of lake sedimentation over nine time periods since 1861. The sedimentation rates have been adjusted to allow for non-catchment derived sediment. (Modified from I. D. L. Foster* et al. *(1985)* Earth Surface Processes and Landforms ***10****, Fig. 9, p.59.)*

area. In a study of a small catchment in Warwickshire, UK, changes in the volumes of sediment deposited in a reservoir in nine periods since 1861 were calculated on the basis of 54 sediment cores dated radiometrically and correlated from magnetic measurements (Fig. 15.14). In this case the source area for the sediment was accurately known (being the present catchment area) and there was also good dating control.

In order to find out about rates of denudation over much longer periods of time we need to apply this approach at much greater temporal and spatial scales. We can, for example, attempt to equate denudation over large areas of continental drainage with the volume of marine sediment deposited offshore along the adjacent continental margin. In effect this is like treating offshore sedimentary basins as enormous, long-lived reservoirs.

The procedure involves four steps. First, the volume and age of sediments offshore must be determined from borehole data and seismic stratigraphy. The major errors here arise from an insufficient coverage of boreholes to give good dating control and the possibility of the addition and removal of sediment by ocean-bottom currents. Secondly, a deduction must be made to allow for the porosity of the sediments and their content of biogenic carbonate deposits (remains of marine organisms). Next, an estimate has to be made of the continental area from which the sediment was derived. This can be a particularly difficult task, especially if we are dealing with drainage areas in existence many tens of millions of years ago. Finally, the mean rate of mechanical denudation is calculated by relating the volume of land-derived sediments over a particular time period to the area of the assumed source region.

In one such investigation of the eastern seaboard of North America it has been estimated that during the Cenozoic about 1000 km^3 of sediment has been produced for each kilometre of the 3000 km of coast from Georgia in the USA northwards to Newfoundland. Deducting 600 km^3 km^{-1} to allow for the contribution of porosity and biogenic carbonate gives 400 km^3 km^{-1} for the volume of land-derived sediment. Assuming that the present crest of the Appalachian Mountains represented the western limit of the sediment source this gives an estimate of 2 km of mechanical denudation over the past 65 Ma for the eastern seaboard of North America at a mean rate of just over 30 m Ma^{-1}. Such estimates are, of course, approximate and do not include chemical denudation for the simple reason that solutes are dispersed throughout the ocean (although they are eventually incorporated either into marine organisms (especially Ca^{2+}) or into deposits on the ocean floor).

15.4.1.2 *Erosion of dated surfaces*

A more direct and location-specific estimation of denudation rates can be made on the basis of the amount of lowering experienced by a surface of known age. This approach can be applied at a range of temporal scales. For instance, it is possible to estimate rates of soil loss from around the roots of trees dated by dendrochronology. Since some species of tree may have long life spans (the 2500 a or more recorded by some bristlecone pines, for example) this technique provides a useful time range. In one study carried out in the Piceance Basin, Colorado, USA, the date when the tree roots were first exposed by erosion was determined by a number of factors including the interpretation of the annual ring growth pattern and the earliest occurrence of reaction wood. Little variation in denudation rates was found between two sites, although a significant difference was found between north-facing slopes with a mean rate of 560 mm ka^{-1} and south-facing slopes with a mean rate of 1180 mm ka^{-1}. The use of the exposure of datable materials can also be applied to human artefacts. Where archaeological remains of known age are present it is sometimes possible to estimate rates of ground lowering, or deposition, since their construction, or abandonment.

For longer time periods it is necessary to use dated surfaces. A classic study using this approach is the investigation by B. P. Ruxton and I. McDougall of the erosion of the Hydrographers Range, a dissected volcanic peak in north-east Papua New Guinea. On the basis of radiometric dating it is known that this composite andesitic volcano last erupted about 650 ka BP and, although the original form of the now eroded upper part of the cone is not known, its likely elevation was about 2000 m. Below about 1000 m the presence of young lavas on interfluves means that the original surface here can be reconstructed fairly accurately. By measuring the cross-sectional area of valleys cut into this surface at a range of altitudes, Ruxton and McDougall were able to estimate the volume of material removed over the past 650 ka. The denudation rate estimated in this case represented the combined effects of mechanical and chemical denudation and was found to increase with increasing local relief and slope gradient (Fig. 15.15).

15.4.1.3 *Fission track and radiometric techniques*

The final approach we need to consider is one which is likely to become increasingly important in studies of long-term denudation rates, namely the exposure of rocks known, on the basis of fission track or radiometric evidence, to have been at a specific depth below the surface at a particular time in the past. We have already outlined the principles of this approach in Section 15.2.1; as pointed out there, although these techniques have been primarily used to infer crustal uplift rates, the direct evidence they provide in fact relates to denudation rates.

It is interesting to see how estimates of denudation rates based on this approach compare with those determined from the volumes of sediment deposited offshore. In a study of the exposure of granite intrusions in northern New England on the northern seaboard of the eastern USA, differences in

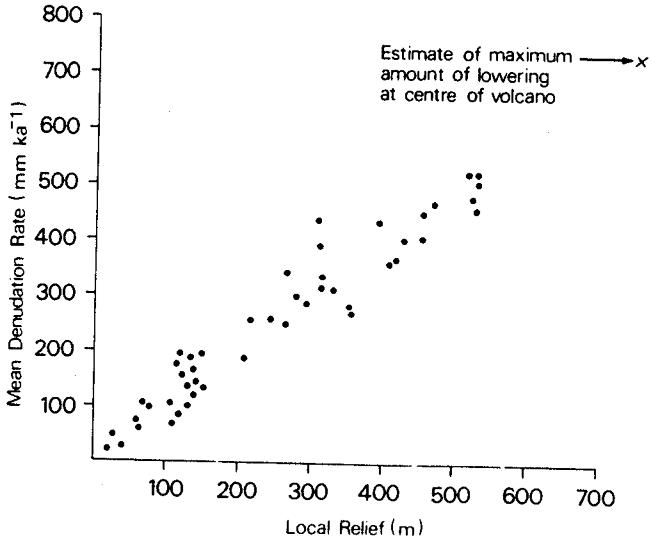

Fig. 15.15 Mean rate of denudation as a function of local relief (difference in elevation between major ridge crests and adjacent valley bottoms) for the Hydrographers Range, Papua New Guinea based on the estimated dissection of dated surfaces. (Modified from R. P. Ruxton and I. McDougall (1967) American Journal of Science ***265****, Fig. 5(A) p. 557.)*

fission track ages between apatite and zircon (the latter having a significantly higher annealing temperature) have been used to estimate their depth and temperature of emplacement. It was calculated that a presently exposed Jurassic intrusion dated at 180 Ma would have solidified at a depth of 5.3–7.6 km and a Cretaceous intrusion (115 Ma) at a depth of 3–3.6 km. These values convert to long-term rates of denudation of 42 to 29 m Ma^{-1} since 180 Ma BP and 31 to 27 m Ma^{-1} since 115 Ma BP. It gives one a sense of confidence that these estimates are broadly in line with the rate of 30 m Ma^{-1} (for mechanical denudation only) for the past 65 Ma estimated on the basis of offshore sediment volumes for the eastern seaboard of North America.

Similarly, comparable results have been established for the Alps where differences in the radiometric ages of biotite and muscovite indicate a mean denudation rate of around 1000 m Ma^{-1}. This contrasts with an estimated mean mechanical denudation rate of 100 m Ma^{-1} for the Rhône Basin in the western Alps based on the sediment volume of the offshore submarine Rhône fan. These results become comparable if it is assumed, as seems likely, that the bulk of the sediment transported by the River Rhône originates from the mountainous Alpine section of its catchment. The highest long-term denudation rate yet recorded appears to be the 10 000 m Ma^{-1} over the past 1 Ma determined from the K–Ar ages of exposed schists in the Southern Alps of New Zealand.

15.4.2 Variations in rates

We have just seen that, on the basis of fission track, radiometric and sediment volume estimates, the range of long-term denudation rates is broadly comparable to that observed for present-day rates of fluvial denudation. We now need to look in more detail at how long-term rates vary over time and space and assess the degree of equivalence between modern and past rates.

One basis for the estimation of long-term global trends in continental denudation is provided by the sediment cores recovered from the ocean basins during the various legs of the Deep Sea Drilling Project (DSDP). This evidence suggests that there have been globally synchronous fluctuations in sedimentation rates (and therefore presumably rates of continental erosion) (Fig. 15.16). High rates during the Middle Eocene (49 to 45 Ma BP) and the Middle Miocene to the present (from 14 BP) greatly exceed those for intervening periods during the Cenozoic. These fluctuations have been interpreted as the result of lower global preci-

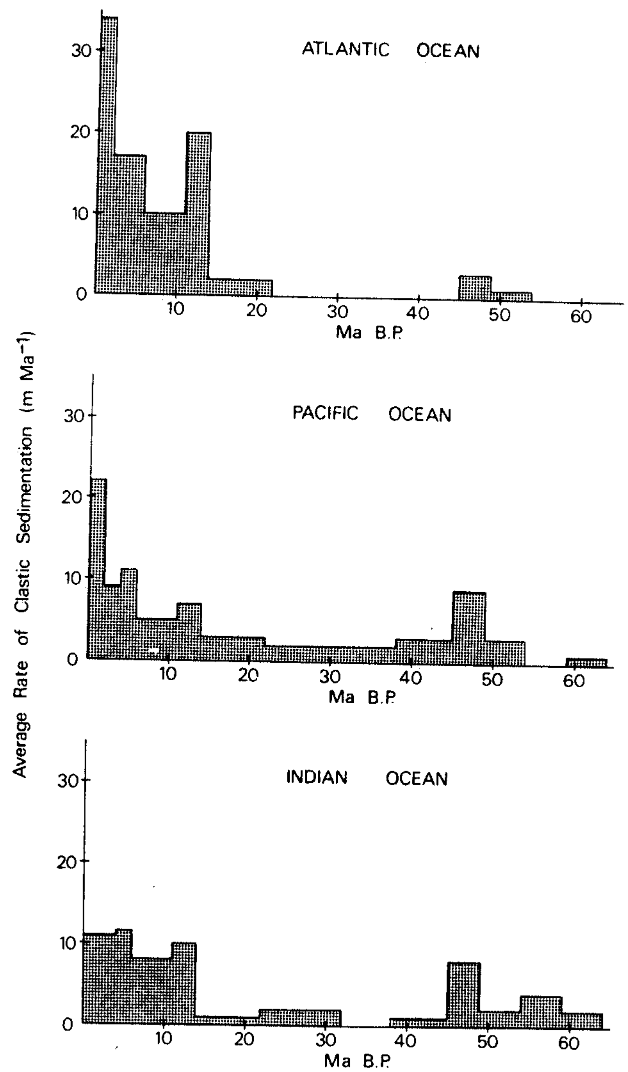

Fig. 15.16 Cenozoic rates of clastic (non-biogenic) sedimentation in the world's ocean basins based on 110 Atlantic, 170 Pacific and 54 Indian Ocean DSDP cores. (Based on data in T. A. Davies et al. *(1977)* Science ***197****, Fig. 1, p. 54.)*

pitation during these periods, although there is evidence that any increased continental denudation during arid phases is likely to be a result of enhanced aeolian erosion (see Section 15.6) rather than greater rates of fluvial sediment transport. However, it is also necessary to consider changes in rates of sediment deposition along continental margins since much of the coarser material removed from the continents is deposited there and fails to reach the deep ocean.

An overriding climatic control has also been invoked by another study examining variations in the rates of deposition of Al_2O_3 recorded in DSDP cores; Al_2O_3 is a good indicator of continental erosion because it is the major element most characteristic of non-biogenic sedimentation. A sixfold increase in the rate of Al_2O_3 accumulation in the northern and central Atlantic and Pacific Oceans over the past 15 Ma indicates a greatly increased rate of denudation during this period, although the increase in the southern Atlantic and Pacific is a more modest 100 per cent. Again, this may reflect primarily increased rates of aeolian denudation since Al_2O_3 could be derived from clay particles in soils deflated from regions experiencing phases of aridity. In a more recent analysis of DSDP cores it has been estimated that the total of denudationally derived sediment and dissolved load delivered to the world's oceans over their present lifetimes of 100–200 Ma is 1.65 $km^3\ a^{-1}$. This suggests a rather low mean rate of continental denudation of 11 m Ma^{-1} (calculated on the basis of the present land area of $148 \times 10^6\ km^2$).

Several estimates of long-term denudation rates in south-eastern Australia based on a variety of methods indicate low rates of landscape modification since the Late Mesozoic (Fig. 15.17). River incision in the upper Lachlan valley of the Murray Basin into radiometrically dated basaltic lavas filling the valley indicates minimum rates of downcutting of 8 m Ma^{-1} for the past 20 Ma and 3–4 m Ma^{-1} for the previous 40 Ma. The rate for the past 20 Ma is broadly confirmed by the rate of sedimentation in the fan at the confluence of the Lachlan with the Murrumbidgee which indicates a rate of 4 m Ma^{-1}. A mean rate of 1–3 m Ma^{-1} is indicated for the whole of the Murray Basin during the Cenozoic, while fission track ages from apatites suggest a denudation rate for the southern part of the East Australian Highlands of 9 m Ma^{-1} over the past 230 Ma or so. Similarly modest rates of denudation of 15–30 m Ma^{-1} are indicated for the past 80–100 Ma following the uplift of the east Australian margin. This evidence is of interest both because of the low rates of denudation that it suggests for the tablelands of south-eastern Australia and for the lack of any evidence of a significant increase in erosion in the Late Cenozoic in parallel with the global trends indicated by the DSDP data (with the possible exception of the Quaternary). This is in spite of evidence indicating increasing aridity and climatic seasonality during this period. The Australian evidence clearly brings into question the global validity of the

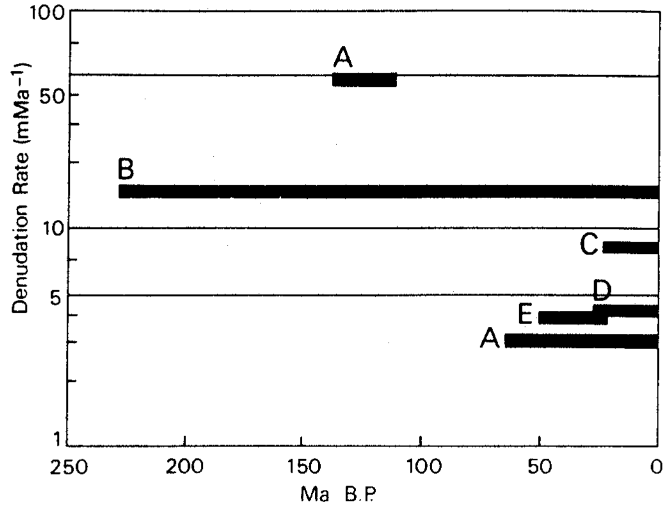

Fig. 15.17 Rates of incision and denudation from the Murray Basin, south-east Australia, based on various methods: (A) basin sedimentation; (B) fission track ages of exposed granitic rocks; (C) river incision into dated basalts, (D) fan sedimentation; (E) river incision between basalts of two ages. (Modified from P. Bishop (1985) Geology *13, Fig. 3, p. 481.)*

trends apparently present in DSDP records and especially the climatic interpretation placed upon them.

What, then, can we say about the present-day relationship between long-term and present-day denudation rates? Although we are living in a rather unusual period of geological time (the Quaternary) characterized by rapid changes in eustatic sea level and major oscillations in global climate, contemporary fluvial denudation rates, perhaps rather surprisingly, do not appear to be significantly different from the long-term average. The estimated present-day global mean of 43 m Ma^{-1} is somewhat higher than the long-term estimate of 11 m Ma^{-1} based on DSDP data, but the latter is very uncertain and is probably a significant underestimate of total continental denudation because it does not adequately incorporate continental margin sedimentation. Certainly, fission track estimates of mean denudation rates of around 30 m Ma^{-1} during the Cenozoic in regions of subdued relief are not incompatible with a global average in excess of 40 m Ma^{-1}. Moreover, the range of present-day denudation rates appears to be very similar to the range evident for the long term, with minima of around 1 m Ma^{-1} and maxima in excess of 5000 m Ma^{-1}. There is evidence that rates in the Late Cenozoic may have increased significantly in some regions as a result of active tectonism or climatic oscillations, and changes in areas of internal and external drainage must also have affected sediment input on to the continental margins.

Although at a global scale there is a broad correspondence between present-day and long-term rates of denudation, anthropogenic effects have clearly become overwhelming in some areas. In Natal in South Africa, for instance, a comparison of modern mechanical denudation rates with long-term rates based on the volume of offshore sediments suggests that the former are between 12 and 22 times the latter. Very high rates of mechanical denudation have arisen over the past few decades as a result of rapid soil erosion arising from poor land management practices associated with excessive population pressures in an environment characterized by steep slopes and periodic intense rainfall.

15.5 Factors controlling fluvial denudation rates

Some of the factors that influence rates of fluvial denudation have already been mentioned in passing and we now consider these factors in more detail. Variations in denudation rates can ultimately be explained by the way erosivity and erodibility interact. As we noted in Section 7.4.4, erosivity is the energy available at the surface to detach and transport regolith; it represents the potential of denudational systems to remove material from drainage basins. As well as the effectiveness of the export of material by streams it also includes the efficacy of other processes involved in sediment transport including rainsplash, sheet flow and rill erosion, and the various mechanisms of mass movement and other denudational processes. In terms of fluvial denudation the important factors affecting erosivity include runoff and the gravitational energy available for transporting material downslope, the latter being related to slope and stream gradients within the basin.

Erodibility is the susceptibility of materials at the surface to transport by denudational processes. This is related to a complex set of factors including mechanical strength, hardness, cohesion and particle size. In addition the rate of solute transport is influenced by the susceptibility of rocks and sediments at, or near, the surface to mechanisms of chemical weathering. The highest rates of denudation will be observed where both erosivity and erodibility are high and, conversely, minimal rates of denudation will occur where both are low. We now look at these factors further.

The strong relationship between relief and denudation rates is evident from the global pattern of denudation (Figs 15.9 and 15.10). It is important here to distinguish between elevation and local relief. The former itself has no direct influence on denudation rates, but the latter is closely related to local slope and thereby to erosivity. As noted in Section 9.2.2, local relief is the difference between the maximum and minimum elevation in an area of specific but limited size. A very high correlation has been established between denudation rate and local relief measured within 20 × 20 km (400 km²) areas across a range of mid-latitude drainage basins (Fig. 15.18). Although it is true that regions at high elevations also tend to have high local relief, especially in orogenic belts, this is far from always the case. Take, for instance, the high plateau of southern Africa, where rather subdued local relief (with equally modest rates of denudation) is found up to 2000 m above sea level.

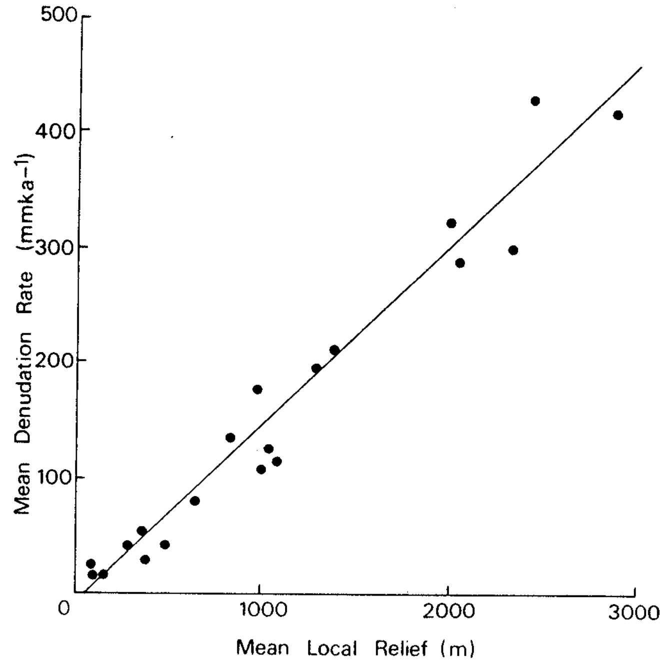

Fig. 15.18 *Relationship between mean local relief and mean denudation rate for 20 mid-latitude drainage basins. (Modified from F. Ahnert (1970)* American Journal of Science ***268****, Fig. 3, p.251.)*

Extremely low rates of denudation are recorded for many of the basins of tropical Africa, northern Eurasia and subarctic North America, in spite of substantial runoff, largely because the local relief in these areas is generally very low. In the Zaire Basin, for instance, gradients are so minimal in the lower part of the catchment that natural lakes have formed.

As we noted in Section 15.3.3 local relief also affects absolute rates of chemical denudation. This is presumably because in regions of high local relief, with steep slopes covered by thin soils or with bare rock surfaces, the potential for chemical weathering is greatest, assuming there is adequate precipitation. This contrasts with areas of low local relief in most climatic environments where the presence of a thick weathering mantle inhibits the movement of water at the weathering front where most of the weathering reactions occur.

Climate has been widely held to have a predominant influence on rates of denudation, although in the light of more recent data it is doubtful whether this view can now be sustained. Several attempts have been made to analyze the relationship between denudation rate and mean annual precipitation (Fig. 15.19). Although there is little agreement in detail between these postulated relationships certain common elements emerge – an initial peak of erosion in semi-arid environments and a progressive increase in denudation above a mean annual precipitation of around 1000 mm. The curve by Ohmori is based on the largest data set and indicates an initial maximum at 350–400 mm mean annual

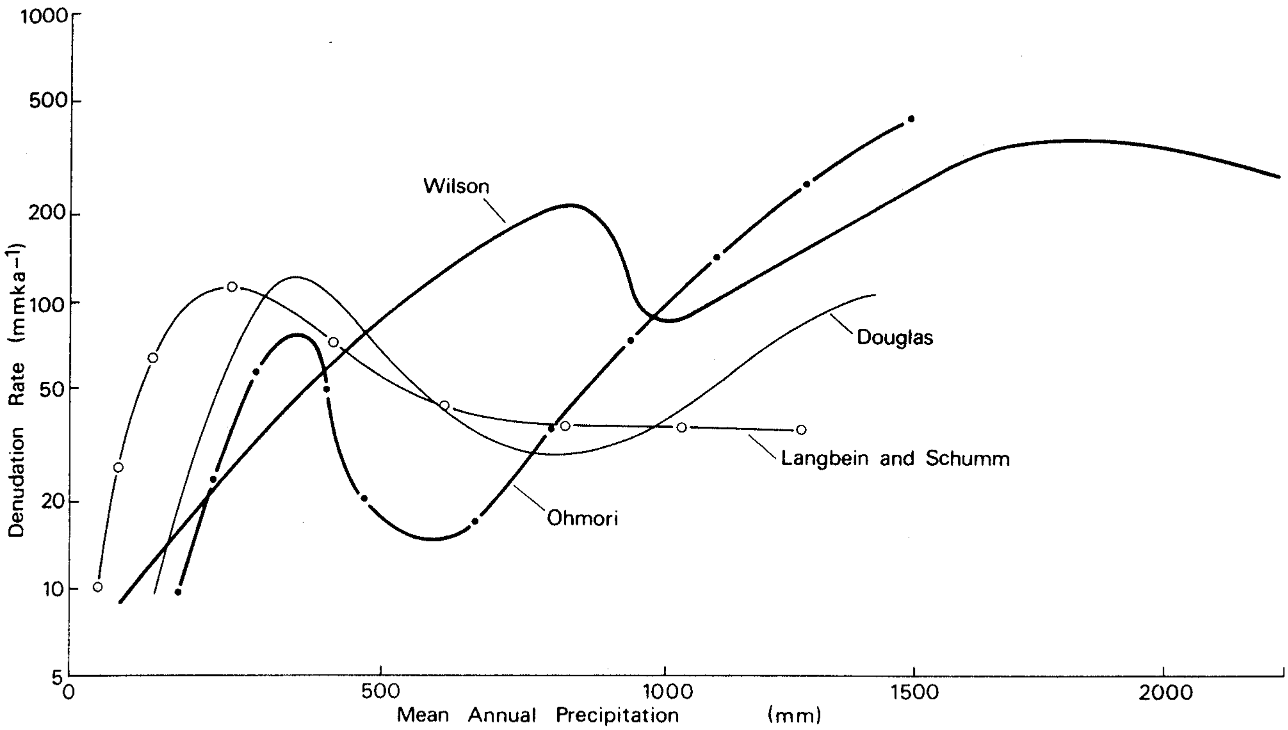

Fig. 15.19 Various estimates of the relationship between (mechanical) denudation rate and mean annual precipitation. (Based on H. Ohmori (1983) Bulletin, Department of Geography, University of Tokyo *No. **15**, Fig. 2, p. 84.)*

precipitation, followed by a minimum at 600 mm and then a progressive increase exceeding the initial maximum at around 1000 mm. This two-maxima curve can be explained by the way the amount of precipitation influences both erosivity and, through the mediation of vegetation, erodibility. The **biomass**, or weight of plant material per unit area, increases with mean annual precipitation up to about 2500 mm a^{-1} in the transition from desert through desert scrub to grassland and forest (Fig. 15.20). The presence of vegetation greatly reduces the erodibility of surface materials, and the abrupt retardation of mechanical denudation rates above a mean annual precipitation of around 400 mm is probably attributable to a marked increase in vegetation cover, as it begins to more than outweigh the effects of increasing runoff. Eventually, however, the maximum protective effect of vegetation is reached and beyond this point increasing precipitation, and therefore runoff, tends to lead to increasing rates of mechanical denudation. The importance of the protective role of vegetation in retarding erosion is dramatically illustrated by the effect of different land uses on denudation rates (Table 15.7).

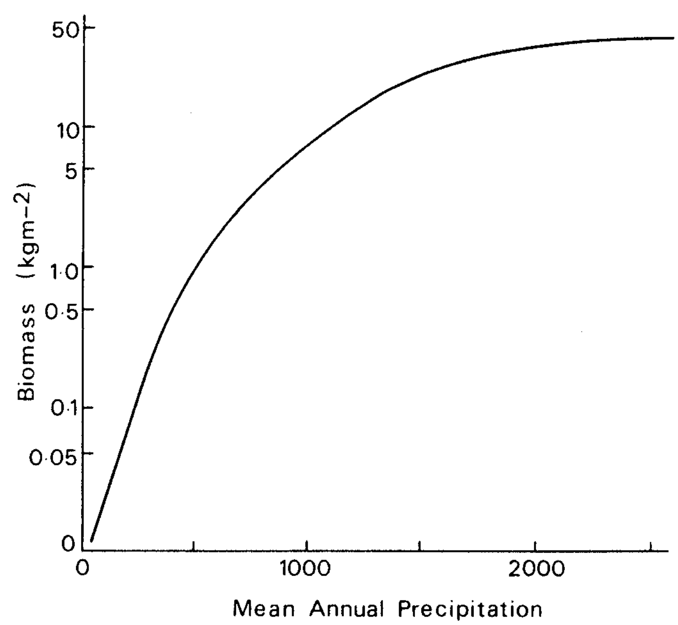

***Fig. 15.20** Generalized relationship between mean annual precipitation and biomass based on various studies. (Based on H. Ohmori (1983)* Bulletin, Department of Geography, University of Tokyo *No. **15**, Fig. 3, p. 86.)*

The relationship between runoff and chemical denudation is rather different as there is a consistent positive correlation between solute load and annual runoff (Fig. 15.21). This is apparently because as precipitation increases there is more water available for chemical reactions in the regolith and solute release, and also greater runoff to transport these solutes. The relationship between chemical denudation and temperature is very weak, however, apparently because the effect of temperature on weathering rates and solute release is overwhelmed by other variables, especially precipitation and local relief.

Table 15.7 Typical rates of erosion for various types of land use in the USA

LAND USE TYPE	SEDIMENT YIELD ($t\ km^{-2}\ a^{-1}$)	DENUDATION RATE ($mm\ ka^{-1}$)
Forest	8.5	4.2
Grassland	85.0	42.5
Cropland	1 700	850
Felled forest	4 250	2 125
Active open-cast mines	17 000	8 500
Construction sites	17 000	8 500

Source: Based on data in Environmental Protection Agency (1973) *Methods for Identifying and Evaluating the Nature and Extent of Nonpoint sources of Pollutants*, Washington, DC.

It has long been recognized that the seasonality of precipitation has a crucial effect on rates of mechanical denudation. This results from both the dramatic increase in sediment-carrying capacity in peak flows and the reduction in the protective role of vegetation during the dry season. The high sediment yields in semi-arid environments may also be influenced by the frequency of high intensity storms, which characterize such regions, as well as the lack of a continuous vegetation cover. In a detailed study in the upper Colorado Basin in the western USA, where the mean annual precipitation ranges from 150 to 1500 mm, rates of mechanical denudation have been found to be negatively related to mean annual runoff but positively related to runoff variability (that is, the frequency of high magnitude flows; Fig. 15.22). The relationship between runoff variability and chemical denudation is, however, negative because as total runoff and, therefore, chemical denudation increases, river discharges become less variable.

A further factor which indirectly influences fluvial denudation rates is drainage basin size. Some studies have indicated that mechanical denudation rates are greater for smaller catchments than larger basins, at least up to a basin size of 2000 km^2 or so. There seem to be a number of reasons for this relationship. Small catchments are commonly in the upper parts of larger basins and they typically have steeper valley-side slopes and channel gradients. Such catchments also have smaller floodplains and less potential for sediment storage. Finally, it is possible for high-intensity storms generating major floods to cover a small basin entirely, whereas such storms will only affect part of a larger basin. Contrasting findings to this inverse relationship between basin area and mechanical denudation rate have come, however, from a study of sediment and solute yields in western Canada. Here the lowest denudation rates occur in the smallest basins (<100 km^2) and the highest in medium-sized catchments (1000–100 000 km^2). The largest

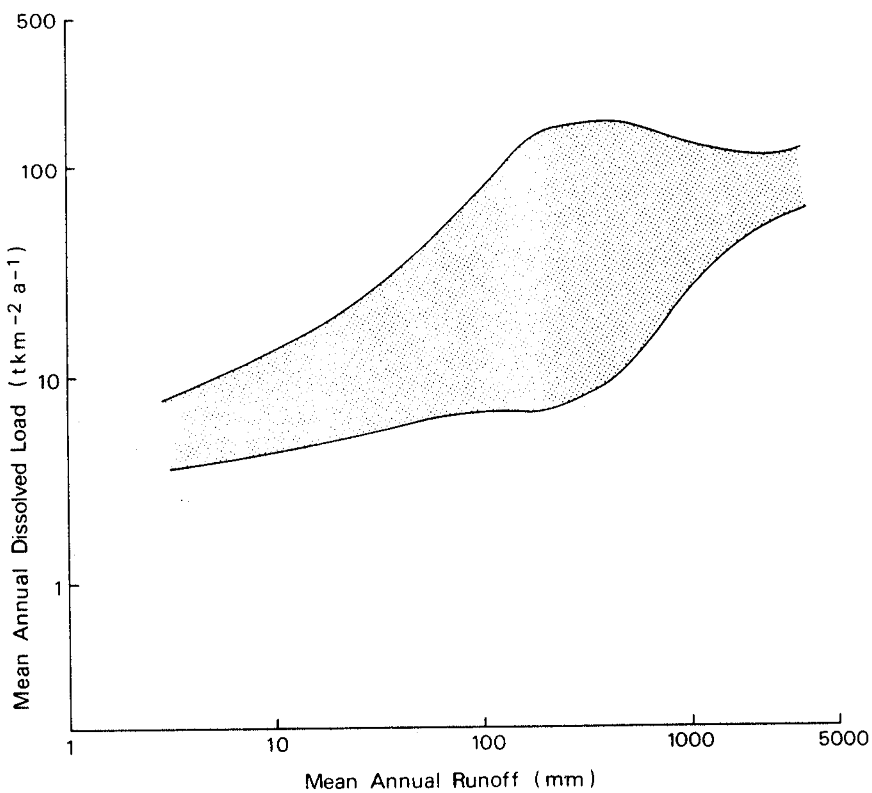

Fig. 15.21 Generalized relationship between mean annual runoff and mean annual solute load for a sample of 496 rivers. The scatter in the relationship is probably due largely to the effects of varying lithology. (Based on D. E. Walling and B. W. Webb (1986) in: S. T. Trudgill (ed.) Solute Processes, *Wiley, Chichester, Fig. 7.3, p. 260.)*

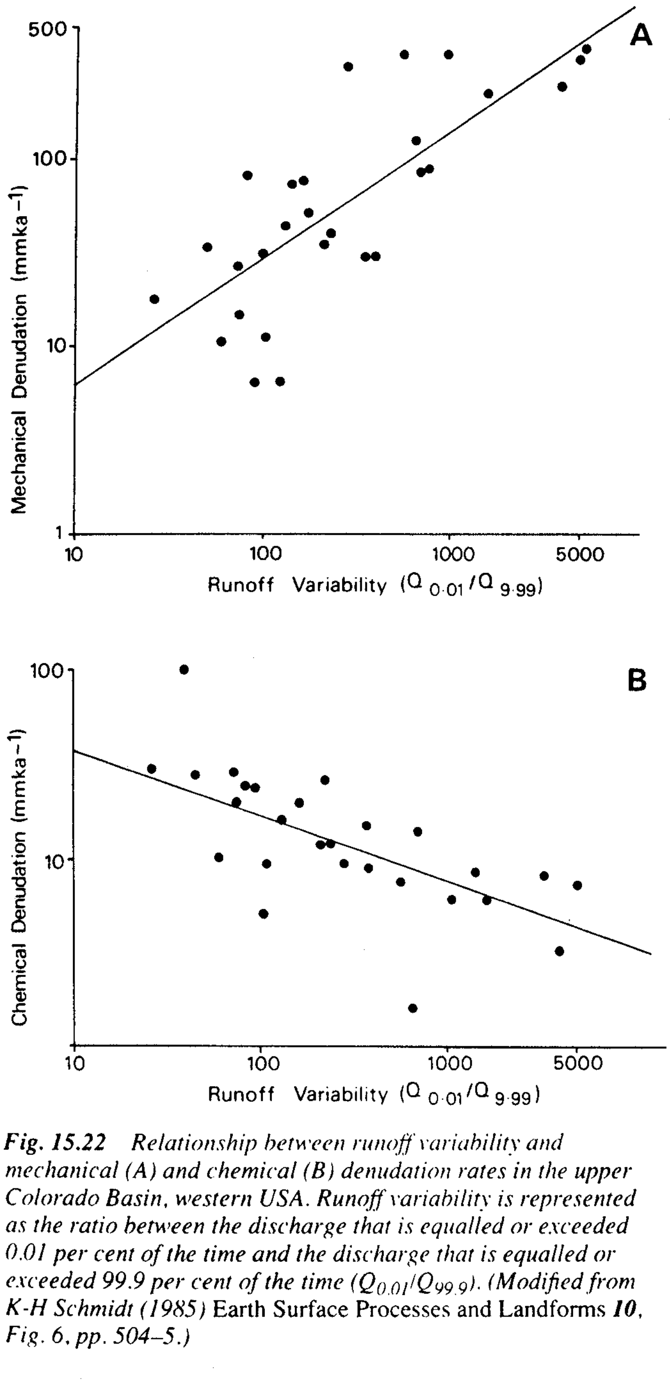

Fig. 15.22 Relationship between runoff variability and mechanical (A) and chemical (B) denudation rates in the upper Colorado Basin, western USA. Runoff variability is represented as the ratio between the discharge that is equalled or exceeded 0.01 per cent of the time and the discharge that is equalled or exceeded 99.9 per cent of the time ($Q_{0.01}/Q_{99.9}$). (Modified from K-H Schmidt (1985) Earth Surface Processes and Landforms ***10**, Fig. 6, pp. 504–5.)*

basins (>100 000 km^2) were found to have intermediate rates.

There are considerable uncertainties as to the dominant factors controlling rates of fluvial denudation, and much research is still needed to test some widely accepted, but not necessarily well supported, generalizations. None the less, it appears that at the global scale, relief and, to a lesser extent, climate are the main determinants of denudation rates (Table 15.8). At the regional and local scale, lithology and the specific factors of erodibility, which determine the supply of sediment and solutes, become more significant.

15.6 Rates of aeolian denudation

Few surveys of global denudation have adequately evaluated the relative importance of aeolian denudation, but it is clear from our discussion of aeolian processes and dust transportation (Chapter 10) that it is potentially significant, at least in arid regions. One reason for this relative neglect is that rates of aeolian denudation are extremely difficult to quantify. Present rates of dust transport from the continents to the oceans can be estimated from samples collected from ships, while individual dust storms can now be monitored by remote sensing techniques either from satellites or by photography from orbiting vehicles such as the space shuttle. Atmospheric transport can also be inferred from the presence of dust derived from remote land areas in the soils of oceanic islands. Such dust may be transported great distances – up to 6500 km from the Sahara to Barbados, 8000 km from the Sahara to Miami, 10 000 km from central Asia to Alaska and 11 000 km from central Asia to the north Pacific islands of Hawaii and Eniwetok (Fig. 15.23).

Estimates of rates of atmospheric dust transport vary enormously from 100 up to 5000 Mt a^{-1}. Assuming all this dust is deposited in the oceans (and therefore lost from the continents) and that the source material density is 2700 kg m^{-3}, this indicates a global mean aeolian denudation rate of between 0.25 and 13 mm ka^{-1}. Any deposition on land would lower these rates, but we must also remember that deflation of dust is confined to a rather small proportion of the total land area, so in these regions the aeolian denudation rate would be much higher than these global average figures. Recent data suggest that the high estimates of several thousand million tonnes per year are, in fact, overestimates, but present-day aeolian denudation rates may still be impressive at the regional scale. It is estimated, for instance, that the rate of dust transport from the Sahara to the Atlantic averages 146 Mt a^{-1}. This implies a sediment yield averaged over the area of the Sahara as a whole of around 16 t km^{-2} a^{-1} which converts to a denudation rate of 6 mm ka^{-1}. However, it is clear that the source of this dust is largely confined to particular regions within the Sahara, especially the Bodele Depression, the alluvial plains of Niger and Chad, southern Mauritania, northern Mali and central southern Algeria, southern Morocco and western Algeria, the southern fringes of the Mediterranean Sea in Libya and Egypt and northern Sudan. If the dust source is limited to these areas then they are probably experiencing a rate of aeolian denudation in excess of 20 mm ka^{-1}, just under half the mean global fluvial denudation rate. Similar estimates of aeolian denudation appear to be valid for east-central Asia, the source of around 20 Mt a^{-1} of fine soil material carried to the north-west Pacific.

Sediment in ocean cores enables past rates of dust deposition to be monitored and both in the north-west Pacific

Table 15.8 Solid and denudational solute load of major rivers and total denudation in relation to climate and relief

CLIMATE AND RELIEF ZONE	SOLID LOAD ($t\ km^{-2}\ a^{-1}$)	DENUDATIONAL SOLUTE LOAD ($t\ km^{-2}\ a^{-1}$)	TOTAL LOAD ($t\ km^{-2}\ a^{-1}$)	TOTAL DENUDATION ($mm\ ka^{-1}$)	TYPICAL SOLUTE LOAD AS % OF TOTAL
Mountainous, high precipitation	200–1500	70–350	250–2000	95–740	10
Mountainous, low precipitation	100–1000	10–60	120–1000	45–370	10
Moderate relief, temperate or tropical climate	40–200	25–60	80–300	30–110	35
Low relief, dry climate	10–100	3–10	15–100	5–35	10
Low relief, temperate climate	20–50	12–50	40–80	15–30	65
Low relief, subarctic climate	1.5–15	5–35	5–40	5–15	80
Low relief, tropical climate	1–10	2–15	4–30	1.5–10	50

Source: Based partly on data in M. Meybeck (1976) *Hydrological Sciences Bulletin* **21**, Table 2, p. 279.

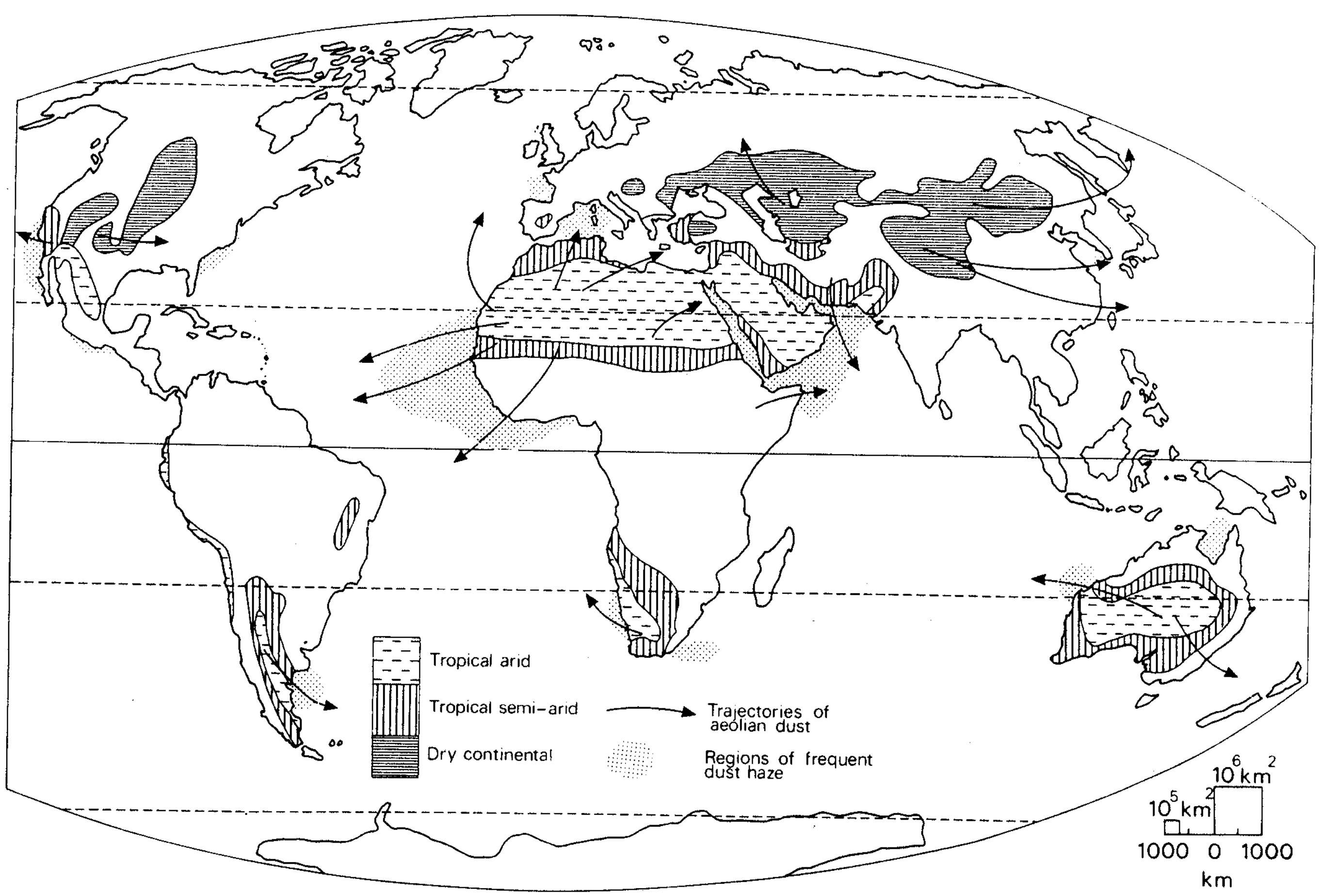

Fig. 15.23 Global distribution of dust storm activity and major dust trajectories. (Modified from N. J. Middleton et al. *(1986) in W. G. Nickling (ed.)* Aeolian Geomorphology. *Allen and Unwin, Boston, Fig. 4, p. 247.)*

and Atlantic Oceans this shows that current rates are representative of the past few thousand years. Taking the ocean core record further back, however, we find some interesting fluctuations in rates of deposition. Around 18 000 a BP rates of accumulation off the coast of West Africa were about twice those of the present, and similar increases at this time have been recorded off coasts adjacent to the Australian, Arabian and Thar Deserts. It is now possible to identify changing rates of aeolian denudation on the basis of atmospheric dust deposition in ocean cores as far back as the Cretaceous. Rates of deposition were apparently low during the Early Cenozoic, but increased after about 25 Ma BP and then accelerated markedly from 7 to 3 Ma BP The most dramatic increase occurred, however, around 2.5 Ma BP accompanying the onset of major glaciation in the northern hemisphere. For much of the Cenozoic the Sahara seems to have remained a major dust source.

Variations in rates of dust deposition in the geological past highlight some of the factors which control rates of aeolian denudation. The dramatic increase in rates at the beginning of the Pleistocene appears to be linked to the establishment of arid – humid climatic cycles at low latitudes at this time roughly in phase with the glacial – interglacial cycles at higher latitudes. This accords with evidence from current dust storm activity which suggests that the most abundant sources of dust are in areas that are presently changing to more arid climatic regimes. There was, for instance, a threefold increase in rates of dust transport out of North Africa during the droughts of the early 1970s and early 1980s. The observation that dust-storm frequency reaches a maximum where mean annual precipitation is around 100–200 mm (Fig. 15.24) is consistent with evidence that water is essential for the weathering processes capable of producing large quantities of the fine sedimentary particles susceptible to deflation. Many hyper-arid areas, by contrast, are relatively 'blown out' with a surface cover of rocky plains or dune fields which yield little of the fine particles necessary for long distance aeolian transport.

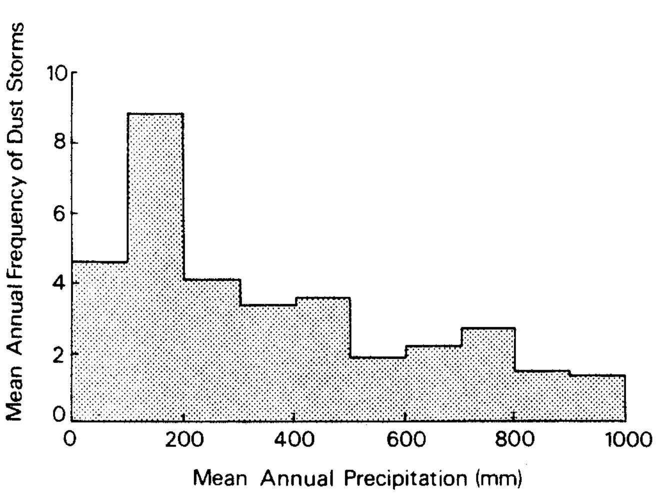

Fig. 15.24 Global mean annual frequency of dust storms in relation to mean annual precipitation. (From A. S. Goudie (1983) Progress in Physical Geography *7, Fig. 11, p. 615.)*

15.7 Rates of glacial denudation

As with aeolian denudation, rates of glacial denudation have received relatively little attention and are similarly difficult to estimate. The methods used include the measurement of the sediment and solute load of glacial meltwater streams, reconstructions of pre-glacial or interglacial land surfaces, the relating of the volume of glacial drift in a specific region to an assumed glacial source area and the estimation of the contribution of glacial debris to marine sediments. There are comparatively few estimates of rates of mechanical denudation for active valley glaciers and these show an extremely broad range from 100 to 5000 mm ka^{-1}. The limited evidence available suggests that rates of solute transport in glacial meltwater streams are relatively high, with a chemical denudation rate of about 20 mm ka^{-1} being calculated for the South Cascade Glacier in the Cascade Range of the north-west USA This is presumably related to the high surface area for chemical reactions provided by the fine particle size of rock flour, and may help to explain the high rates of chemical denudation estimated for partially glaciated mountain belts.

There is considerable uncertainty as to the rates of denudation associated with continental ice sheets. This is illustrated by the debate over the depth of erosion accomplished by the Pleistocene Laurentide ice sheet in eastern North America. One view is that the glacial scouring of this region has been largely confined to the exhumation of a weathering front formed under pre-glacial conditions and has been limited to an average of a few tens of metres at most. In support of this idea is the absence of the quantities of glacial till to be expected if deep erosion had occurred. On the other hand, it has been pointed out that much of the glacial debris produced could have been transported offshore by meltwater streams. Recent estimates of marine deposition rates along the eastern margin of North America do indeed suggest a significant input of glacially derived material during the Pleistocene. Although the precise source region for this sediment is impossible to define, it has been estimated that the volume deposited offshore implies a minimum mean depth of erosion of 120 m for the Laurentide region with a possible maximum of 200 m. This gives a mean denudation rate of between 48 and 80 m Ma^{-1} throughout the Pleistocene, and this is clearly rather higher than would be expected of fluvial denudation for such an area of subdued relief.

Rates of sediment production from 'permanent' ice masses, such as the Antarctic ice sheet, are very low and it appears that significant erosion is only accomplished by periodic ice sheets such as those of the northern hemisphere in the

Pleistocene. Whatever the general picture it is clear that erosion under even periodic ice sheets can be highly selective. This is illustrated in the Buchan area of north-east Scotland where, in spite of successive ice sheet advances across the region, are still preserved deep weathering mantles of pre-Pleistocene age.

15.8 Comparison of rates of uplift and denudation

Few topics in geomorphology have received a more confused analysis than the relationship between rates of uplift and denudation and the implication this has for the way landscapes evolve. We will leave the question of landscape evolution until Chapter 18, but we will conclude this chapter by considering relative rates of uplift and denudation.

In an influential paper published in 1963, S. A. Schumm asserted that modern rates of orogenic uplift at around 7500 mm ka^{-1} are some eight times greater than average maximum denudation rates. This conclusion has subsequently been widely accepted and has been interpreted as meaning that the assumption that episodic phases of rapid uplift punctuate the continuous but more leisurely progress of erosion is broadly correct. Such an assumption has important implications for the understanding of landscape evolution so we must ask, is it really true? Schumm was working with a much smaller database than that now available and his estimates of maximum rates of denudation are unduly modest. We have seen, in fact, that maximum rates of both orogenic uplift and denudation lie between 5000 and 10 000 mm ka^{-1}. Rates of glacio-isostatic rebound can, of course, be even higher than this but such vertical displacements can only amount to a total of a few hundreds of metres.

The evidence now available on rates of crustal uplift and denudation in orogenic belts indicate that the interaction between uplift and denudation has an inherent tendency towards a steady state where both are roughly equal in rate. During the uplift of a mountain range the denudation rate along its crest is initially slower than the rate of crustal uplift. This is because as we have seen, the denudation rate is largely a function of local relief, and this in turn is related to the degree of fluvial dissection. As this begins at the margins of an uplifted mountain range and progresses towards its crest by headward erosion, the summit region is relatively unaffected by denudation in these early stages of uplift. Consequently the elevation of the crest of the range increases, but as fluvial dissection works towards the crest the rate of denudation increases until eventually rates of uplift and denudation are roughly equal. In spite of further crustal uplift, in part promoted by isostatic rebound as a consequence of continuing denudational unloading, the mountain range cannot become any higher as long as the long-term rate of crustal uplift remains roughly constant and there are no significant changes in exogenic conditions. This equivalence of rates of crustal uplift and denudation in young mountain ranges is not fortuitous but simply reflects the achievement of a steady state, and the time taken to attain the steady state elevation is largely a function of the rate of uplift.

But how high can a mountain range become before this steady state is achieved? This seems to depend largely on the width of the mountain range, since the broader the range the longer it takes for fluvial dissection to reach the crestal regions, and therefore the longer the time during which the crustal uplift rate on the divide exceeds the rate of denudation. Consequently, it is no surprise to observe that the

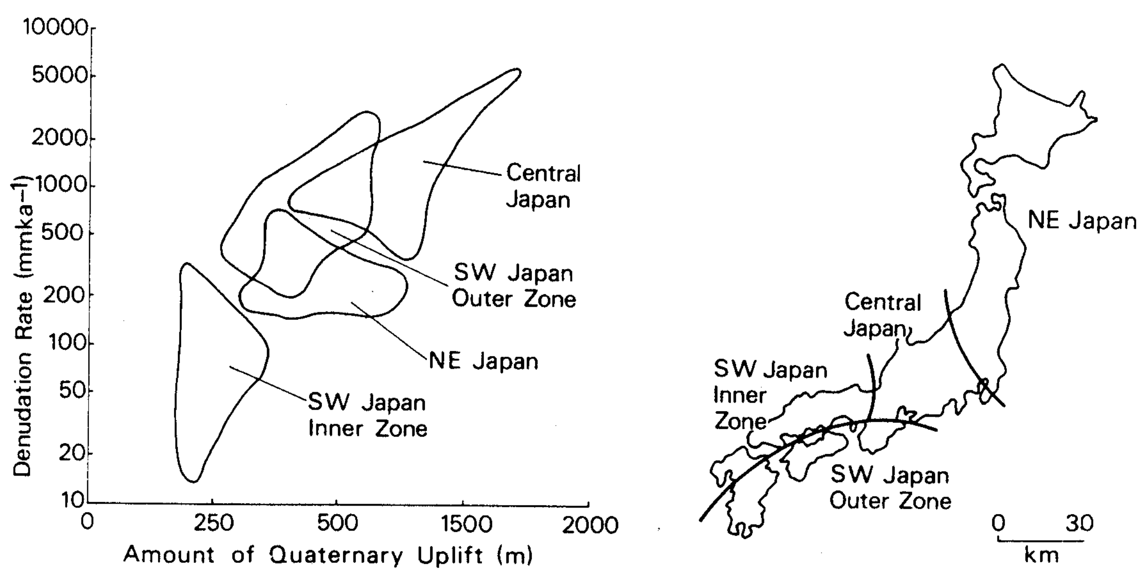

Fig. 15.25 Relationship between rates of Quaternary uplift and rates of present denudation in Japan. (Based on T. Yoshikawa (1985) in A. Pitty (ed.) Themes in Geomorphology. *Croom Helm, London, Fig. 12.3, p. 203, and Fig. 12.4, p. 204.)*

crest of the relatively narrow Southern Alps of New Zealand, which are only some 80 km across, has an average height of about 3000 m whereas the Himalayas, which have a width of around 350 km, have an average crestal elevation of around 7500 m.

The relationships between rates of crustal uplift and denudation in orogenic belts can be instructively examined by looking at New Zealand and Japan, both regions for which relatively good data are available. In Japan the central zone with crustal uplift rates of about 2200 mm ka^{-1} has attained a steady state and the outer zone nearly so, whereas in the north-east and south-west zone crustal uplift still exceeds denudation (Fig. 15.25). It appears that in the central and south-west zones the rapid rates of crustal uplift have prompted a rapid response in the rate of denudation, whereas in the other two zones the slower rates of crustal uplift have not yet led to a fully compensating increase in denudation rates. A similar picture is revealed in South Island, New Zealand, where rates of crustal uplift increase rapidly towards the west as the zone of oblique convergence marked by the Southern Alps is approached (see Section 3.5.2; Fig. 15.26). The crest of the Southern Alps seems to be in a steady state with very high rates of both crustal uplift and denudation in excess of 5000 mm ka^{-1}. On the eastern flanks of the Southern Alps, however, in areas such as eastern Otago, rates of crustal uplift are considerably lower (100–300 mm ka^{-1}), but still well in excess of rates of denudation at around 70 mm ka^{-1}. Here, then, there has apparently been insufficient time for a steady state between crustal uplift and denudation to be established.

So far we have limited our discussion of the relationship between crustal uplift and denudation rates to the situation occuring in orogenic belts. What of those regions experiencing epeirogenic uplift? Excluding areas of glacio-isostatic rebound, long-term rates of epeirogenic crustal uplift appear to be of the order of 100 mm ka^{-1} (in spite of possible short-term rates an order of magnitude higher than this). However, the low local relief of many areas experiencing

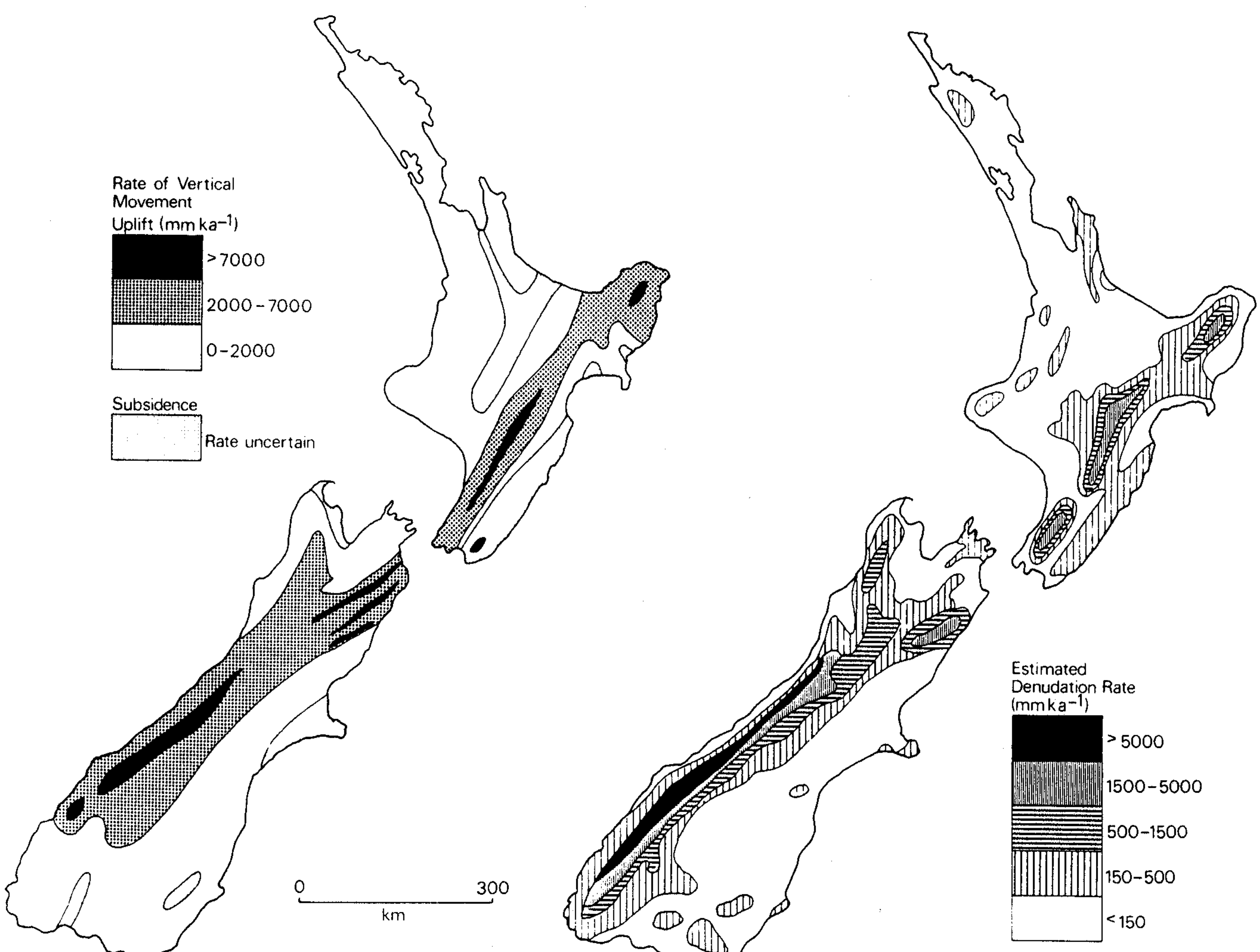

Fig. 15.26 Comparison of rates of crustal uplift and denudation in New Zealand. (Modified from M. J. Selby, (1982) Hillslope Materials and Processes. *Oxford University Press, Oxford, Fig. 11.11, p. 237.)*

epeirogenic uplift, such as the regions of broad continental warping in Africa and Australia, mean that denudation rates in such areas are often even lower. Consequently, there is little prospect of a steady state being achieved in these cases, a point which has interesting implications for models of landscape development as we will see in Chapter 18.

Further reading

There is an enormous range of literature on rates of uplift and denudation, but fortunately much of this is summarized in reviews. Most of the material on uplift rates is not set specifically in a geomorphic context, but the importance of such information for understanding landform development cannot be overemphasized. It is, however, important to be aware of the confusion between surface uplift and crustal uplift evident in much of the literature on uplift rates.

The most useful general assessment of crustal movements and the techniques used to monitor them, especially over the short to medium term, is the book by Vita-Finzi (1986). The ambiguities inherent in the measurement of present-day deformation using relevelling surveys are discussed by Brown *et al.* (1980), while Adams (1984) looks at a range of evidence for recent crustal movements along the northern Pacific coast of the USA. Schaer *et al.* (1975) compare modern and past rates of uplift in the Swiss Alps, while various aspects of the literature on rates of tectonic uplift are briefly reviewed in my short discussion of neotectonics and landform genesis (Summerfield, 1987). Rates of glacio-isostatic uplift are considered by Nikonov (1980). Clark and Jäger (1969) discuss the application of radiometric dating to estimating crustal uplift rates, while Gleadow and Fitzgerald (1987), Moore *et al.* (1986) and Parrish (1983) apply fission track dating techniques. Both present-day and long-term rates of uplift in orogenic belts are considered by Gansser (1983).

There have been several recent assessments of estimates of rates of mechanical and chemical denudation based on sediment and solute load data for rivers. Sediment yields are dealt with by Milliman and Meade (1983) and Walling and Webb (1983a), while the solute loads of rivers are considered at a predominantly local and regional scale by Walling and Webb (1986) and at a largely global scale by Walling and Webb (1983b). These wide-ranging reviews also consider the problems of estimating sediment and solute loads and of inferring patterns of denudation from them. On this theme Meybeck (1983) and Cryer (1986) deal in detail with the question of atmospheric inputs in river solutes, while Dunne (1978) provides an excellent investigation of chemical denudation rates carefully excluding non-denudational components. Turning to sediment loads, Douglas (1967) assesses the importance of anthropogenic factors, while Trimble (1977, 1983) and Meade (1982) address the important effect that the disequilibrium between sediment supply and transport can have on estimates of denudation rates. Saunders and Young (1983) and Young and Saunders (1986) provide useful overviews of the relative importance of different denudational processes under various climates, and Meybeck (1976) and Clayton and Megaham (1986) provide, respectively, a general and specific assessment of the relative importance of mechanical and chemical denudation.

At the longer time scale, Carrara and Carroll (1979) provide a case study estimating erosion rates from tree root exposure, while Foster *et al.* (1985) demonstrate the variability in sediment transport evident in lake sediment records over periods of a few decades. Assessments of long-term global rates of sediment supply to the oceans include those by Davies *et al.* (1977) (updated by Worsley and Davies, 1979), Donnelly (1982) and Howell and Murray (1986). Matthews (1975) uses the offshore sedimentary record to examine Cenozoic rates of erosion in eastern North America, while Ruxton and McDougall (1967) document the rate of fluvial dissection of the Hydrographers Range in Papua New Guinea. The use of fission track and radiometric ages to determine long-term average denudation rates is illustrated in the papers already mentioned on the estimation of crustal uplift rates, and by Doherty and Lyons (1980). Bishop (1985) convincingly demonstrates the predominance of low rates of denudation throughout the Cenozoic in south-eastern Australia, while Young (1983) briefly examines the evidence for, and implications of, low denudation rates on a broader basis. At a more local scale Martin (1987) compares long-term sedimentation rates off the Natal coast with contemporary sediment yields. The factors determining denudation rates are examined briefly in the reviews of solid and solute load data mentioned above, but more detailed evaluations are provided by Wilson (1973) on climatic controls, Ahnert (1970) on the local relief factor, Schmidt (1985) on the importance of runoff variability, Schumm (1963) and Slaymaker (1987) on drainage basin area and Ohmori (1983) and Trimble (1988) on the role of vegetation. The influential paper by Langbein and Schumm (1958) is still worth a look as long as its limited applicability is appreciated.

The role of dust transportation is considered by Goudie (1983) and Middleton *et al.* (1986), while Lever and McCave (1983) use evidence from deep sea cores to trace the variations in aeolian deposition in the Atlantic since the Cretaceous. Rates of sediment transport in glacial streams are examined by Gurnell (1987), and the high solute concentrations typical of glacial meltwater are considered by Collins (1983). The debate concerning rates of ice sheet erosion is illustrated through the contributions of White (1972), Sugden (1976) and Bell and Laine (1985), while Hall and Sugden (1987) provide specific evidence of the inefficiency, at least locally, of ice sheets as denudational agents.

On the question of the relative rates of uplift and denudation it is appropriate to begin with Schumm's influential assessment (Schumm, 1963) and then to look at the excellent discussions by Adams (1985) on the Southern Alps and Yoshikawa (1985) on Japan. The factors determining the elevation of a mountain range are considered by Ahnert (1984).

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