modification has been extremely slow, apparently involving an average of less than 100 m extension over the past 5 Ma. In the eastern Mojave Desert 60 basalt flows up to 4 Ma old partly bury pediment surfaces, and comparison of the relative positions of pediments covered by dated basalts and the modern pediment surfaces indicates the degree of modification that has occurred over this time interval. The conclusion is the amount of modification has been minimal, with downwearing predominating over backwearing.

Although the exhumation model of pediment formation, along with its requirement for a previously humid climate, seems to have widespread applicability there appear to be situations where it cannot apply because there is compelling evidence of prolonged aridity. One such case is provided by the pediments flanking the domed inselbergs, or **bornhardts**, of the Namib Desert in southern Africa (Fig. 14.3). Here the cold Benguela Current offshore in conjunction with the dominant sub-tropical high pressure system has created a hyper-arid coastal desert climate which has probably prevailed since at least the mid-Miocene. Although the present rate of slope retreat seems to be extremely slow there are no remnants of an earlier phase of significant rock weathering to support the idea of exhumation. Such pediment and inselberg forms represent one of the major enigmas of geomorphology.

14.2 The significance of climatic change

The presence of landforms, such as certain kinds of pediments, in climatic environments in which they could not have developed obviously raises the question of climatic change in landform interpretation. The state of the atmosphere varies from day to day and even hour to hour, and it is average weather conditions together with their variability that characterize the climate of a region. But climate also fluctuates at all time scales from decades to tens of millions of years. Whether a climatic change is of geomorphic significance depends on its magnitude and duration, and on the properties of the landform concerned. The larger or more resistant a landform, the longer, in general, it takes to adjust to a change in climate. The 'sensitivity' of different landscapes to climatic fluctuations is an important but complex question which affects the way we conceive long-term landscape development and we examine it in detail in Chapter 18 (see Section 18.2.3).

An important distinction is between 'active' landforms which seem to be in 'equilibrium' with the prevailing climatic environment, and inactive or relict landforms which could not have been formed under the current climatic regime. As will have been evident from our discussion of the pediment problem, this distinction is not always easy to make. Although we saw in Section 1.5.3 that it is possible to relate broad associations of geomorphic processes to major climatic zones, the overlap between categories is significant since most processes operate in most environments at least to some degree. In many cases we do not yet know enough about the development of specific landforms to be able to say with any confidence which are being actively formed in a particular climatic environment. The problem of distinguishing between active and relict forms can be particularly acute in those environments where even 'active' geomorphic processes operate sporadically. This is the case in many arid and semi-arid environments where the irregular patterns of precipitation mean that fluvial activity is very limited in duration, but, none the less, often very important in shaping the landscape. The question here is how long does a landform have to be inactive for it to be described as relict?

The past two decades have seen a quantum leap in our understanding of climatic change largely as a result of the exploration of the ocean floor, which contains a relatively complete and undisturbed climatic record, and of the development of new dating procedures and techniques of palaeoenvironmental reconstruction. Our knowledge of climatic fluctuations on the continents, nevertheless, is still severely constrained by a lack of palaeoclimatic indicators capable of yielding datable material. In fact landforms themselves have been extensively used to reconstruct changes in temperature, precipitation and wind intensity and direction on the continents. The use of this kind of evidence actually raises a problem for geomorphologists concerned with interpreting the landscape in terms of climatic change because in many cases the climate changes themselves have been interpreted largely from landform evidence. Clearly there is a danger of circular arguments here and, if possible, we must use independent information on climatic change.

The consequences of climatic change for landform genesis can, very broadly, be divided into those effects related *primarily* to temperature changes, and those related *primarily* to changes in precipitation. The growth and decay of ice sheets is clearly influenced significantly by long-term changes in temperature, whereas surface runoff and fluvial processes, along with the level of aeolian activity, are highly responsive to precipitation changes. We must be careful, however, in making broad generalizations. For instance, although frequently resulting from a temperature fall, glacier growth can also occur as a result of an increase in precipitation when temperatures are stable, or even increase. Moreover, an increase in surface runoff can arise from a decrease in temperature (and hence rates of evapotranspiration) as well as an increase in precipitation. Nevertheless, it is convenient to separate the discussion of the expansion and contraction of glacial and periglacial morphoclimatic regimes, which are largely a response to temperature fluctuations, from the predominant impact of precipitation changes on fluvial and aeolian systems. Before we do this, however, it is necessary to review briefly our present understanding of the history of global climatic change.

14.3 The record of climatic change

A wide range of techniques are used to reconstruct past climates and they are employed on diverse types of evidence. For long-term climatic change information is acquired from the changing distributions of plants and animals determined from fossils, and from rock types which are attributable to particular climatic regimes. Evaporites and aeolian sandstones, for instance, are indicative of arid environments, whereas tillites demonstrate glacial conditions. Approaches to reconstructing more recent climatic change include the analysis of pollen (**palynology**) and landforms themselves.

One technique, however, has done more than any other over the past two decades or so to revolutionize our understanding of climatic change. In the late 1940s it was pointed out that the ratio of the two stable isotopes of oxygen, ^{16}O and ^{18}O, precipitated in carbonate would vary with water temperature. This idea was applied in 1955 by C. Emiliani to the analysis of calcium carbonate ($CaCO_3$) shells secreted by various sea creatures. The technique of **oxygen isotope analysis**, as it was termed, was most often applied to foraminifera, single-celled organisms mostly between 0.1 and 0.3 mm across. Emiliani found that the $^{18}O/^{16}O$ ratio varied in foraminifera shells recovered from ocean cores recording sedimentation over the past few hundred thousand years. Both planktonic (surface and near-surface) and benthic (bottom-dwelling) forms of foraminifera occur so it was possible to estimate temperature changes on both the sea surface and in the ocean depths. The changes in oxygen isotope ratios were dated using the palaeomagnetic time scale in combination with an assumed constant rate of sedimentation.

It was subsequently appreciated that the $^{18}O/^{16}O$ ratio of the sea water itself would also vary as the quantity of land ice changed with fluctuations in global temperatures. Such changes would be reflected in the foraminifera and would arise because the lower atomic mass of ^{16}O means that it is preferentially evaporated from the oceans. During glacials a proportion of this ^{16}O-enriched water would be locked up in ice sheets thus leaving the oceans relatively enriched in ^{18}O. It is difficult to separate these two effects – one indicating changes in ocean temperatures, the other fluctuations in the volume of ice sheets – and this has led to divergent views of climatic history, especially for the past 40 Ma.

14.3.1 The Cretaceous to Neogene record

There are probably few areas in the world where landforms or weathering deposits can be regarded as having survived more or less intact for more than the past 100 Ma, although it has been suggested that even more ancient forms exist in northern Australia. It is appropriate, therefore, to begin our brief survey of climatic change with the Cretaceous.

The climate of the Late Cretaceous – 100 to 65 Ma BP – is generally thought to have been warm and equable with a very much less marked temperature gradient with latitude compared with the present. Polar regions probably experienced a climate rather similar to mid-latitude and subtropical climates of the present day, although the amount of solar radiation received and the seasonality of climate would have been similar to equivalent latitudes today. However this widely accepted ice-free Cretaceous interpretation has been challenged by researchers who have interpreted large exotic blocks found within Early Cretaceous mudstones in Australia as **dropstones**, that is, debris dropped from floating ice. This evidence indicates that ice was present at sea level in high latitudes during the Cretaceous. Similar interpretations have been applied to exotic blocks in strata of a variety of ages and it has been argued that these indicate, contrary to conventional wisdom, that the Earth may only rarely have been ice free in the past. These interpretations have yet to be confirmed by other evidence, and meanwhile it is perhaps best to adopt the generally accepted ice-free Cretaceous model. Towards the end of the Cretaceous there is good evidence for a slight cooling, but this trend was reversed during the Paleocene and Eocene when a warm phase saw tropical soil formation extending to 45° latitude in both hemispheres and tropical species living in southern England and western Greenland.

The major point of dispute in the climatic record for the past 50 Ma is when the Antarctic ice sheet became established. One interpretation of the oxygen isotope data suggests that it first appeared about 15 Ma ago, but if $^{18}O/^{16}O$ ratio variations are caused predominantly by ice volume changes then a date of around 35 Ma is indicated (Fig. 14.4). This earlier date is supported by geomorphic and sedimentological evidence, including glacially abraded grains in offshore sediments, which indicate the first appearance of sea-level glaciation in East Antarctica at this time. By about 26 Ma ago the East Antarctic ice sheet was apparently more extensive than it is at the present day. An ice sheet in the Early Cenozoic also helps to account for the record of sea-level change during this period which is difficult to explain without the changes in ocean water volume generated by periodic fluctuations in ice volume (see Section 17.5.4).

The causes of the major climatic changes involved in the establishment of the Antarctic ice sheet are beyond the scope of this book, but continental drift and the drastic modifications of ocean currents that it generated were certainly important factors. Recall that the supercontinents of Laurasia and Gondwana broke up progressively throughout the Cretaceous (see Figure 2.16). In the northern hemisphere a temperate climate persisted until the mid-Miocene, but significant cooling can be traced back to at least 10 Ma ago and the Arctic Ocean became ice-covered by 5 Ma BP. Although some uncertainty remains, it appears that ice sheets became established in the northern hemisphere around 3.2 Ma BP. This date is indicated by the appearance of ice-rafted debris in the North Atlantic and North Pacific in sediments of this age.

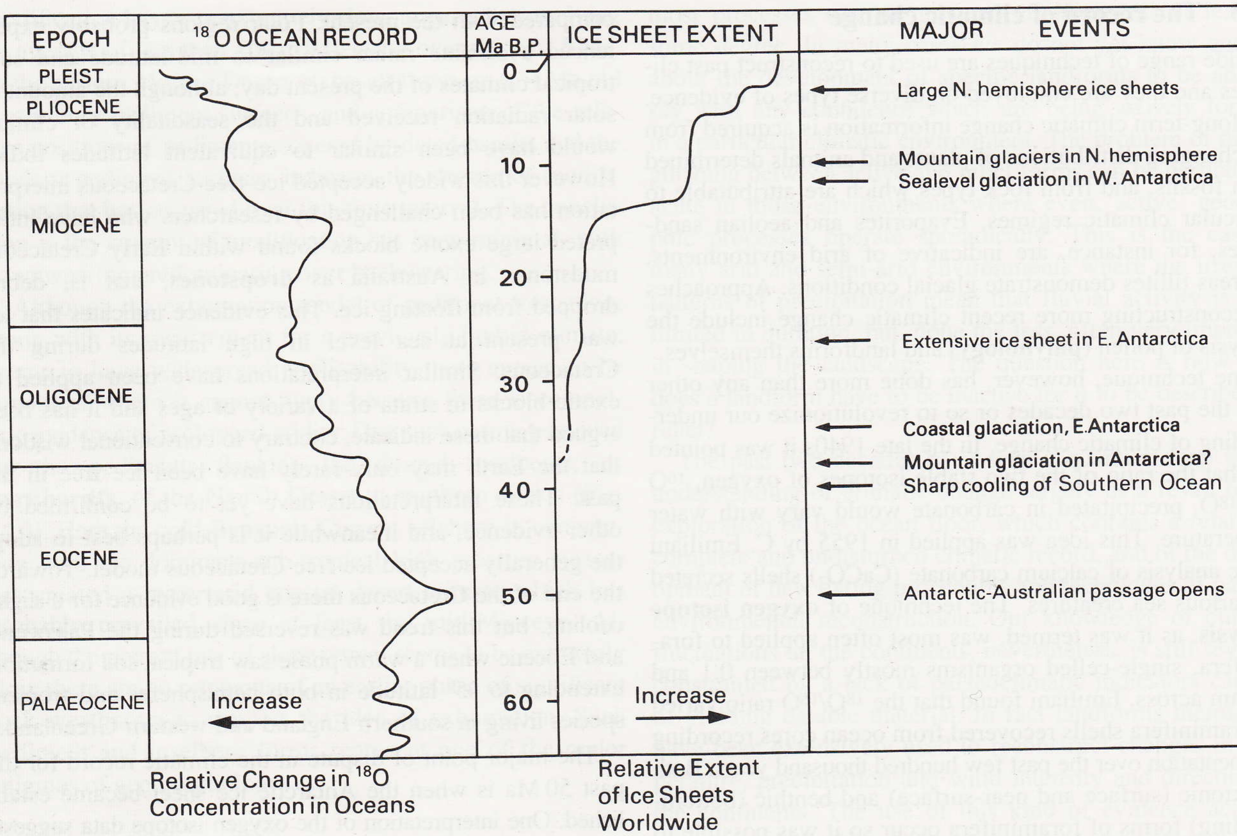

Fig. 14.4 *History of the development of glaciation during the Cenozoic. Note that the variations in ice sheet extent are speculative due to lack of data, and that the dates at which glacial activity began at different localities is also subject to revision in the light of new information. (Based partly on D. E. Sugden (1987) in M. J. Clark, K. J. Gregory and A. M. Gurnell (eds)* Horizons in Physical Geography. *Macmillan, Basingstoke, Fig. 3.2.3, p. 217.)*

14.3.2 The Quaternary record

By 3 to 2.5 Ma BP large-scale oscillations in temperature reflecting glacial–interglacial cycles had become established, with a regular 100 ka cycle occurring over the past 700 ka. Many researchers attribute these regular climatic fluctuations largely to the effects of slight variations in the position and orientation of the Earth with respect to the Sun which affect the receipt of solar radiation; this is commonly known as the Milankovitch mechanism for glacial–interglacial cycles. Other research, however, has indicated that other factors, such as the organic productivity of the oceans and the concentration of CO_2 in the atmosphere, may also be important. The oxygen isotope data indicate that over the past 3 Ma the Earth has experienced a glacial climate for a significant majority of the time, with glacial episodes of around 80–100 ka duration alternating with 10 ka–20 ka interglacial interludes, at least over the past 700 ka. Significant changes in the $^{18}O/^{16}O$ ratio are now indicated by **oxygen isotope stage numbers** and these provide a good guide to the major changes in ice sheet volume and, indirectly, global temperatures (Fig. 14.5). Key age markers for this ocean core oxygen isotope record have been provided by the palaeomagnetic time scale and there have also been attempts to link the ocean record with continental palaeoclimatic data. The most successful of these correlations has been achieved with loess sequences (see Section 10.3.7). The longest record is provided by the loess sequences of China which seem to give a continuous climatic record for the past 2.3 Ma. Enhanced rates of deflation and aeolian dust transport occurred during periods of semi-arid climate coinciding with glacial advances in the northern hemisphere. A correlation has now been made between the past 500 ka of a Chinese loess sequence with the ^{18}O record of a deep-sea core in the north-west Pacific and this has greatly improved the dating of the loess sequence (Fig. 14.6).

A major contribution to the understanding of the nature and magnitude of Quaternary climatic change has come from the CLIMAP (Climate/Long-Range Investigation Mapping and Prediction) project which attempted to document the global environment at the last glacial maximum 18 ka BP.

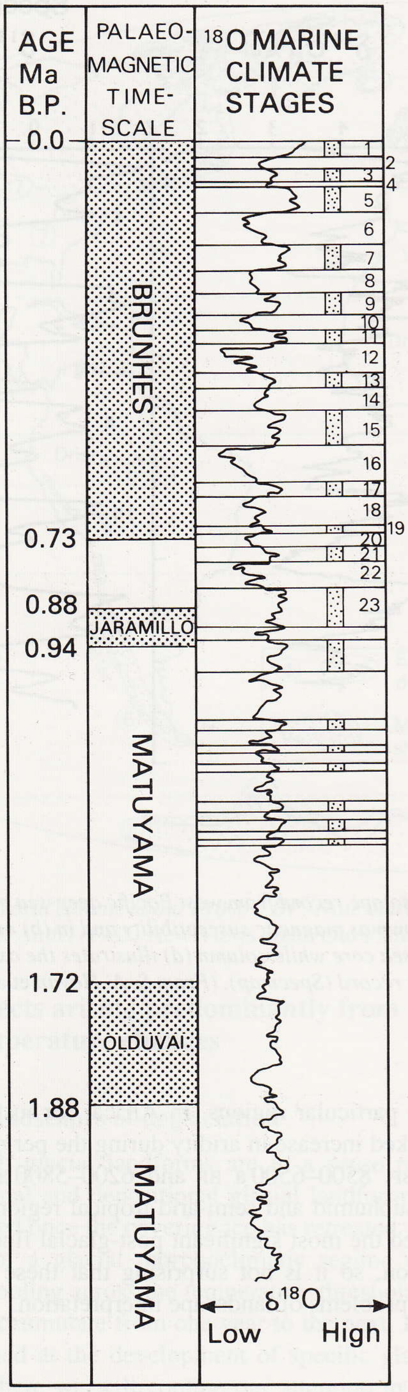

Fig. 14.5 Oxygen isotope record for the about the past 2 Ma from the Pacific deep-sea core V28-239 analyzed by N. J. Shackleton and N. D. Opdyke correlated with the palaeomagnetic time scale of Berggren et al. *The oxygen isotope stage numbers are shown, the odd-numbered shaded stages indicating relatively warm interludes. (Modified from W.B. Harland* et al. *(1982)* A Geologic Time Scale. *Cambridge University Press, Cambridge, Chart 2.17, p. 42.)*

Although concerned with just the most recent major glacial advance, the environmental changes identified were probably mirrored by earlier glacials of similar magnitude. Large ice sheets blanketed a significant proportion of the land area of the northern hemisphere and some 18 per cent of the Earth's surface was ice-covered (Fig. 14.7). In eastern North America the Laurentide ice sheet extended from the Rocky Mountains to the Atlantic Ocean and from the Arctic Ocean to the latitude of New York. In the ice free area to the south large lakes formed, such as those of the Great Basin of the western USA, in response to the supply of meltwater, and to lower evaporation as a result of lower temperatures.

Until the 1960s it was widely thought that the humid tropics had been little affected by the glacial–interglacial fluctuations of the higher latitudes. New data from a range of sources, including palynological studies of sediment cores and work on changing lake levels, have now shown that, during the latter part of the last glacial at least, much of the humid tropics were probably cooler and drier than at present (Fig. 14.8). Although the evidence is sparse in some regions, during the period around the last glacial maximum at about 18 000 a BP the tropical rain forests seem to have contracted to a smaller area than they presently cover. There was a compensating increase in the extent of savanna vegetation, indicating a significant reduction in precipitation. According to some interpretations of the sedimentological evidence up to half of the land area within 30° of the equator was covered by active ergs at this time, and lake levels, especially in the northern hemisphere, were generally very low in the period 14 000–21 000 a BP. The evidence for the magnitude of temperature depression is less certain with a greater temperature decrease being inferred from the lowering of the altitude of vegetation and glacier limits than is indicated by information from deep-sea cores.

It is in fact not a surprise to find evidence of tropical aridity coinciding with glacial episodes. We would expect a reduction in precipitation to occur as a result of reduced evaporation in response to lower ocean surface temperatures and a decrease in the area of the oceans as consequence of a global fall in sea level. Exceptions to this glacial aridity trend did occur though, with some regions such as North Africa being more humid (Fig. 14.8). This was due to the equatorward deflection of rain-bearing westerly weather systems by the large high pressure cells established over the extensive northern hemisphere ice sheets.

Climatic changes have continued since the end of the last glacial around 10 000 a BP but their magnitude has been much attenuated. Although mean global temperature does not appear to have varied by more than about 2 °C over the past 8000 a or so, evidence from fluctuating lake levels indicates that the tropics were generally much wetter in the period 12 000 to 5000 a BP than they were during the preceding 10 000 a, or have been over the past 5000 a. This humid episode was, however, punctuated by much drier

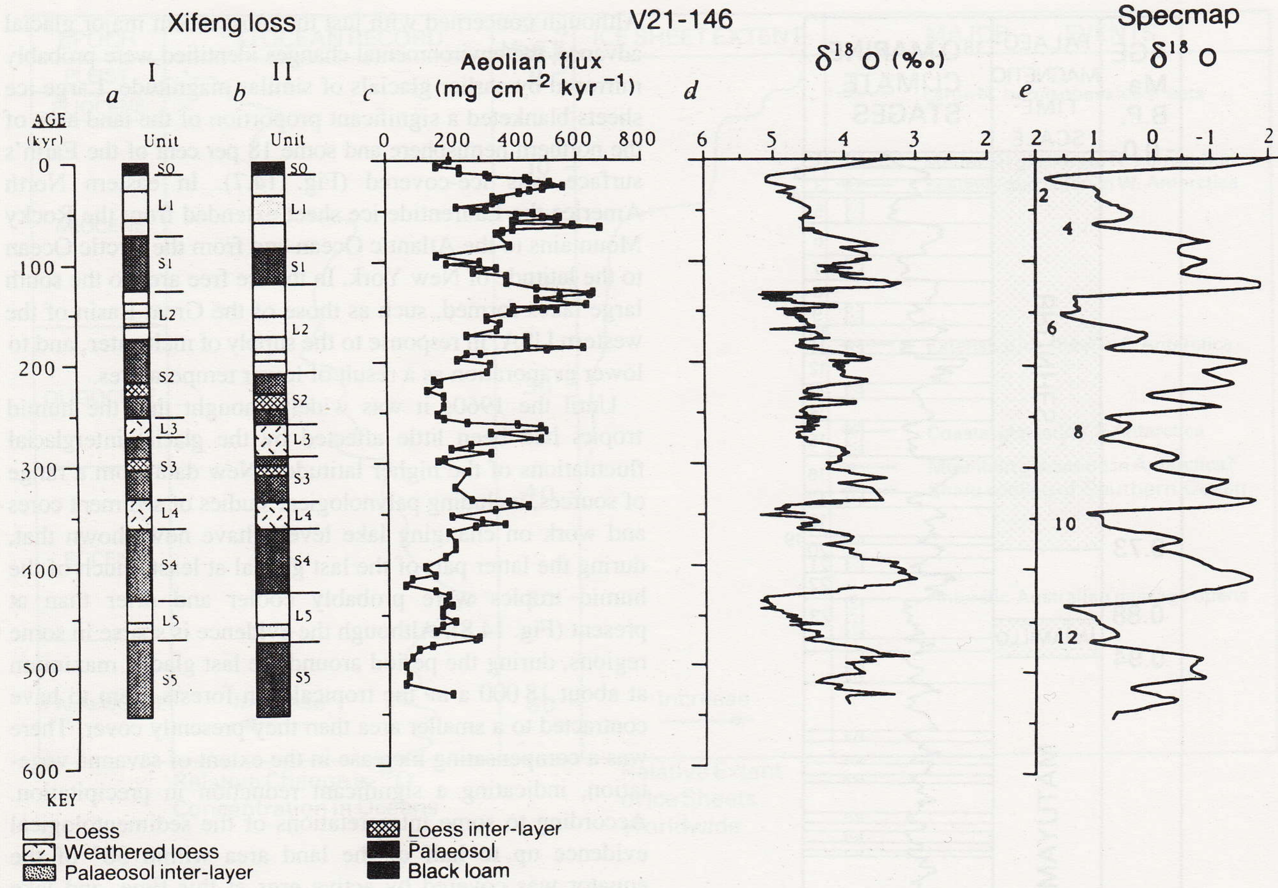

Fig. 14.6 Correlation of the loess sequence at Xifeng, China, with the oxygen isotope record from west Pacific deep-sea core V21–146. Dating of the loess sequence in column (a) is based on a technique known as magnetic susceptibility and in (b) on correlation with the aeolian dust flux. Column (c) shows the aeolian flux record of the deep-sea core while column (d) illustrates the oxygen isotope record. Column (e) shows the global standard chronology for the oxygen isotope record (Specmap). (From S. A. Hovan et al. *(1989)* Nature *340, Fig. 3, p. 298.)*

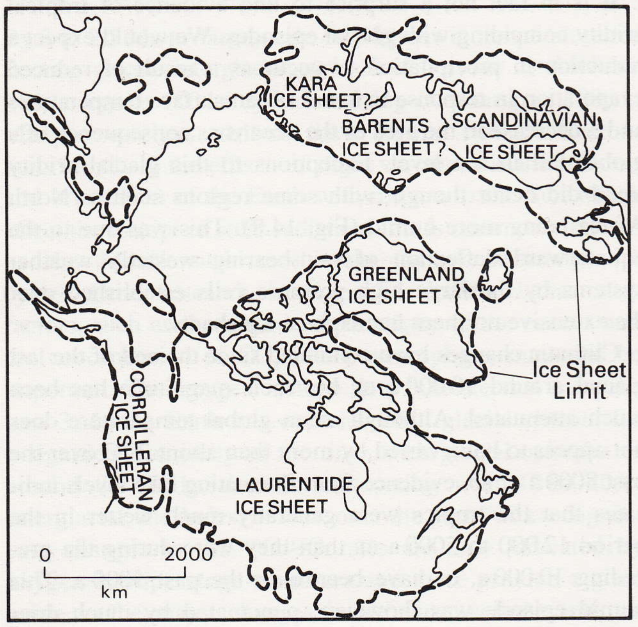

Fig. 14.7 Probable extent of the northern hemisphere ice sheets at the last glacial maximum 18 000 a BP (Based largely on G. H. Denton and T. J. Hughes, (1981) The Last Great Ice Sheets. *Wiley, New York, p.viii.)*

periods in particular regions. In Africa, for instance, there was a marked increase in aridity during the periods 11 000–10 000 a BP, 8500–6500 a BP and 6200–5800 a BP. It has been the subhumid and semi-arid tropical regions that have experienced the most significant post-glacial fluctuations in precipitation, so it is not surprising that these areas pose particular problems of landscape interpretation.

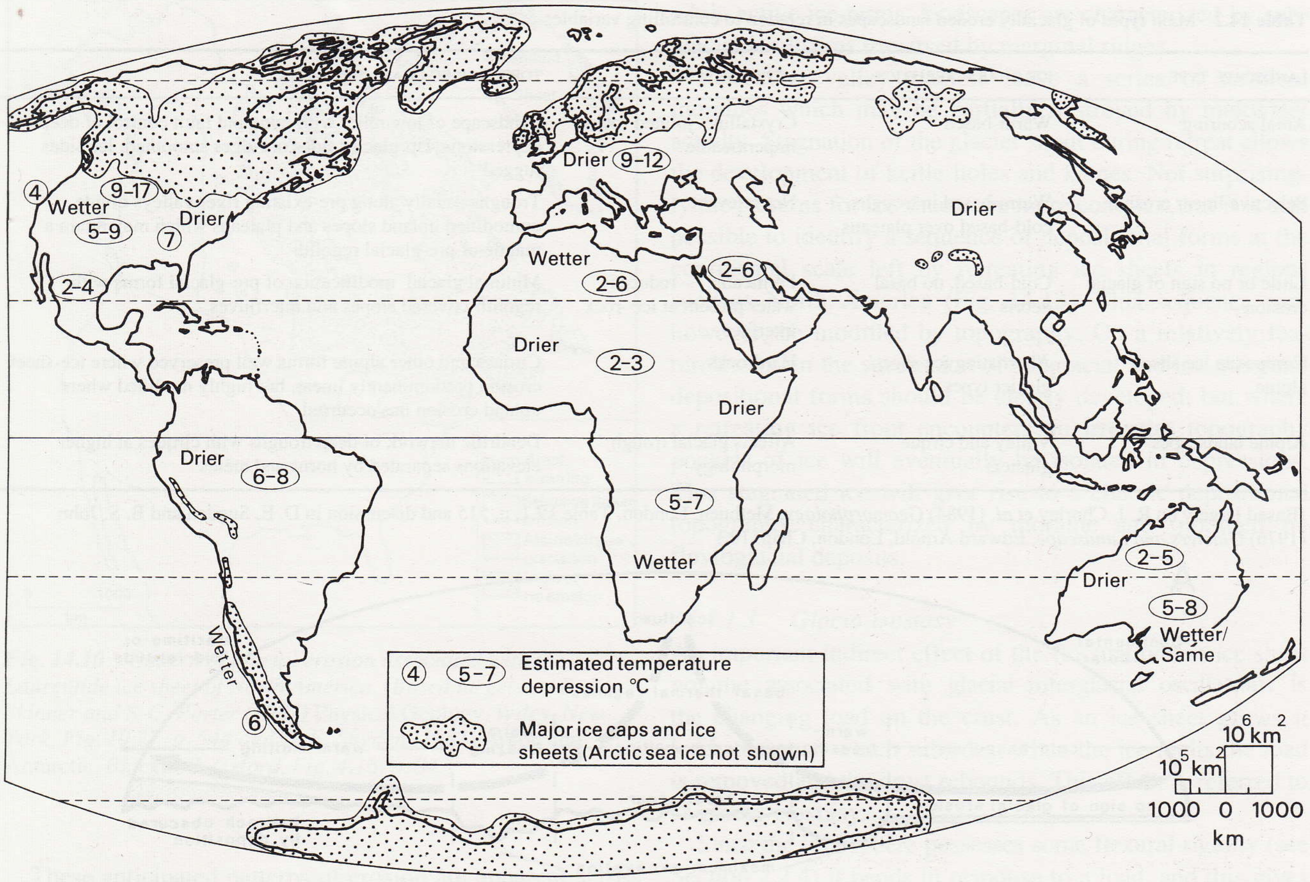

Fig. 14.8 World climate about 18 000 a BP. Note that the palaeoclimatic reconstructions for some areas are uncertain. (Based partly on T. H. Van Andel (1985) New Views on an Old Planet. *Cambridge University Press, Fig. 4.4, p. 61 and various other sources.)*

14.4 Effects arising predominantly from temperature changes

14.4.1 Landscapes of deglaciation

Virtually all glacial landforms are in a sense relict since both erosional and depositional glacial landforms are only fully exposed once the covering ice has retreated. Similarly, the onset of a glacial morphoclimatic regime requires a sustained cooling across the temperature threshold at which snow can accumulate from one year to the next. In Chapter 11 we looked at the development of specific glacial landforms, but here we will confine our attention to the broad nature of landscape modification that resulted from the expansion of the Late Cenozoic ice sheets into the mid-latitudes (Fig. 14.7) and the extension of upland ice caps and mountain glaciers to lower elevations.

14.4.1.1 Relict landscapes of glacial erosion

Mountain, or alpine, glaciation is characterized by a deepening of valleys by glacial erosion and active physical weathering on valley sides. Glacial troughs develop generally along the course of pre-existing river valleys (Fig. 11.16); bedrock resistance influences glacial trough form with deeper and narrower valleys occurring in resistant lithologies. In regions of modest elevation, landscape modification during glacials may be confined to the formation of cirques. In the mid-latitudes of the northern hemisphere cirques are preferentially developed on north-east slopes. Snow accumulates here because of the combination of shade from summer insolation with protection from the predominant westerly winds. In the southern hemisphere south-east slopes are favoured for similar reasons.

The nature of landscape modification by ice sheets is controlled by ice sheet behaviour, the bedrock geology and the pre-existing form of the landscape (Table 14.2 and Fig. 14.9). The crucial factor seems to be the state of the basal ice. Active erosion can occur below warm-based ice and extensive areal scouring is accomplished where the pressure melting point is attained, normally beneath the centre of ice sheets and around the margins in lower latitudes. This broad pattern can be modified by the subglacial topography since upland regions will be more likely to be covered by thinner, cold-based ice. Only very limited erosion is achieved beneath the cold-based polar flanks of continental ice sheets.