Chapter 28

Cenozoic Climates in Deserts

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Introduction

Deserts are superb repositories of geological, geomorphic and archaeological evidence. The very aridity to which they owe their existence has enabled them to preserve a remarkably good record of past depositional and erosional events. The fossil river valleys of the Sahara, the great salt lakes of Australia, China, and Patagonia, the dissected volcanic mountains of the Arabian peninsula and the Afar Desert – all are legacies of former tectonic, volcanic, and climatic episodes which ultimately gave rise to the deserts we see today. Each desert reflects its own individual geological inheritance and geomorphic history; each is unique in its assemblage of landforms; each ideally deserves detailed and separate study in its own right (Pesce 1968, McKee 1978, Rognon 1989).

Two contrasting themes permeate the study of desert landscapes: longevity and change. In many parts of the desert world, morphogenesis is virtually inactive, and relief is presently being conserved. In other more limited areas, recurrent dust storms and ever shifting dunes convey an impression of a dynamic and changing landscape. This latter impression is reinforced by the presence of now vegetated and stable dunefields located well beyond the present-day confines of active desert dunes, as well as by the occurrence of relict river and lake basins deep within the desert – silent witness to previously wetter climates. The paradox here is that deserts can be simultaneously both young and old, conserving as well as destroying relief, morphologically stable as well as geomorphically dynamic. In the deserts of Western Australia, for instance, late Quaternary dunefields adjoin valleys that have changed very little since the final separation of Australia from Antarctica $\sim$50 Ma ago. The frequent juxtaposition of very old elements of the landscape with others that are very young reflects the polygenetic nature of desert landscapes (Mabbutt 1977, Frostick and Reid 1987) .

The aim of this chapter is to examine the role of past tectonic, volcanic, and climatic events in shaping the landscapes of our major deserts, focusing upon the Cenozoic legacy in particular, while noting that many aspects of desert geomorphology cannot be fully understood without an appreciation of much older tectonic and sedimentary events, many of them extending well back into Phanerozoic times, and some even into the Precambrian. We begin with a summary of the causes of present-day aridity, followed by a discussion of the problems and assumptions involved in reconstructing Cenozoic climatic changes in deserts. We then consider the geomorphic history of the North African and Australian deserts, and conclude with an evaluation of the impact of Cenozoic climatic fluctuations upon the evolution of desert landscapes worldwide.

Causes of Aridity

We will now examine some of the causes of present-day aridity and some of the factors which control the distribution of our existing deserts. Deserts are regions of rare and unreliable rainfall. They are not restricted to any particular latitude, but are especially extensive

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astride the two tropics, in latitudes characterized by more or less permanent high pressure cells and hot dry subsiding air. The two polar deserts are under the influence of semi-permanent anticyclones and of cold dry subsiding air. As air subsides and becomes compressed, it also becomes warmer, so that its relative humidity is decreased even though the absolute amount of water vapour held in desert air may be substantial and may become evident in the cold hours before dawn in the form of the evanescent desert dew. The anticyclonic hot deserts result from the global Hadley circulation, and have very little to do with the regional distribution of land and sea. Oceans in strictly tropical and high polar latitudes also receive minimal precipitation and are simply the oceanic extensions of the terrestrial deserts.

A second and very common cause of aridity on land is distance inland. Except for the equatorial zone, precipitation usually decreases rapidly away from the coast. The great temperate deserts of the continental interior of Asia are a prime example of such deserts, and the effects of continentality reinforce those of latitude in the hot tropical deserts of North Africa, Arabia, and Australia.

Two additional factors accentuate the aridity resulting from latitude and increasing distance from the nearest source of moist maritime air. One is the rain-shadow effect, and the other is the presence offshore of cold upwelling water or a cold ocean current. When warm moist maritime air reaches the land it frequently encounters mountain ranges, as in the case of the Andes, the Rockies, the Himalayas, the Ethiopian uplands, and the Eastern Highlands of Australia. The moist air rises and is cooled adiabatically, rapidly attaining vapour saturation, and shedding its precipitated water as rain or snow – hence such local names as Sierra Nevada and Snowy Mountains. If the coastal ranges are reasonably elevated relative to the interior, as is true of the Americas and Australia, there will be a pronounced rain-shadow effect inland of the ranges, with desiccation of the previously moist air accentuated as it flows downhill, becoming warmer and drier in the process. In extreme cases, the air may shed its moisture on mountains 2000–4000 m high, before descending to valleys lying close to or even below sea level, as in the Afar Depression and the Dead Sea Rift.

Deserts such as the Atacama and the Namib are flanked offshore by cold upwelling water. In fact, the western borders of all the Trade Wind or tropical deserts are washed by the cool waters of ocean currents generated by the gyres which flow clockwise in the Northern Hemisphere and anticlockwise in the Southern Hemisphere. The result is predictable: cool moist air from the ocean reaches land which is warmer than the adjacent cool ocean, so that the relative humidity of the air is reduced, and its capacity to absorb moisture rather than to shed it is increased. Such air masses therefore usually function as a desiccator rather than as a welcome source of moisture, leaving the plants and animals dependent on coastal fog for their survival. Erkowit, situated high in the Red Sea Hills of the eastern Sudan, is a good example of a mist oasis, receiving much of its precipitation from fog blowing off the Red Sea in winter while all around it is sweltering lowland desert.

Aridity should not be confused with desertification. Within the timescale of the last two million years the deserts have occupied essentially their present locations on the globe (Fig. 28.1). They are where they are for sound and relatively immutable geographical reasons. A combination of at least five major factors accounts for the distribution of modern deserts and for their low and erratic precipitation (Williams 2002a). These are the prevalence of dry subsiding air over the deserts (itself linked to latitude and to global atmospheric circulation), a vast land area, low inland relief and high coastal ranges, and the presence of cool ocean water close offshore. An additional factor is the presence aloft of a subtropical jetstream, the existence and course of which are partly controlled by the presence or absence of extensive areas of high elevation, such as the Tibetan Plateau immediately north of the Himalayas, which exerts a strong influence on the easterly jetstream which flows from Tibet across the Arabian peninsula towards Somalia (Rognon and Williams 1977, Flohn 1980), accentuating the aridity in those regions.

The previous discussion now requires some qualification, elaboration, and amendment. All of the 'relatively immutable' geographical controls over aridity alluded to in the previous paragraph are true only for the very late Cenozoic. Prior to that, lithospheric plate movements created a different and constantly changing distribution of land and sea, of warm and cold ocean currents, and of high and low terrestrial relief. We will discuss the global climatic repercussions of Cenozoic plate tectonic movements in the final section of this chapter.

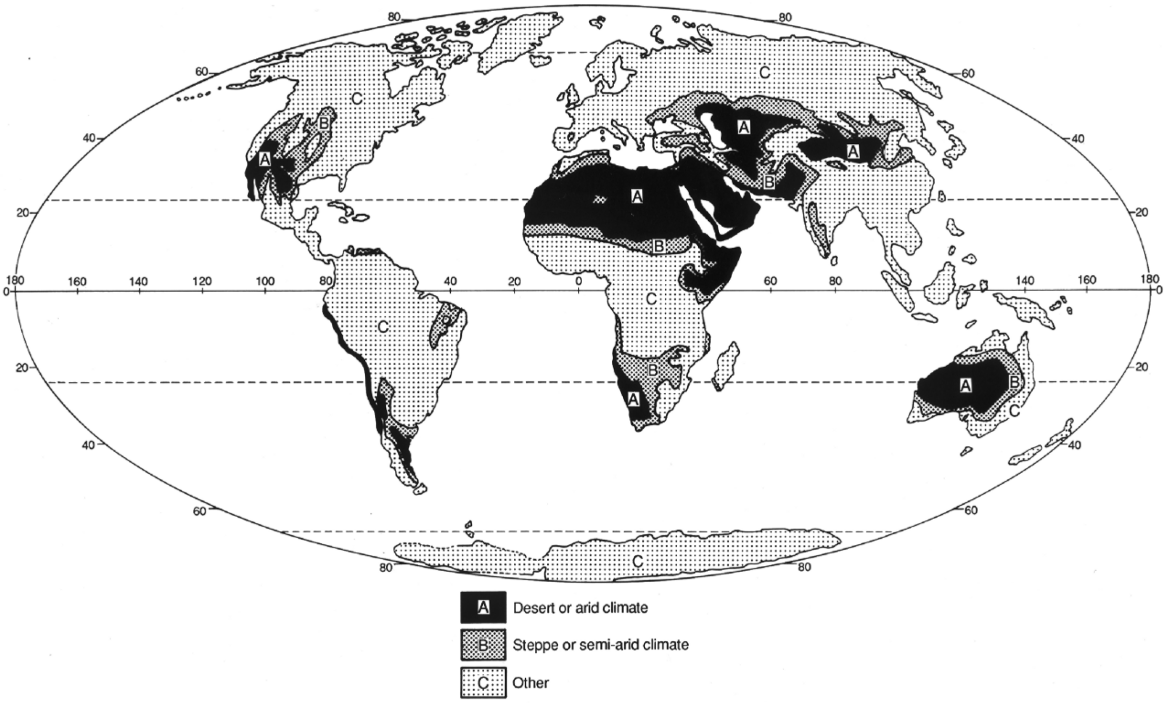

Fig. 28.1 The semi-arid and arid regions of the world

Even within the limited time frame of the Quaternary (c. 1.8 Ma), the sphere of influence of the deserts has waxed and waned from times of maximum expansion, such as during the very late Pleistocene $\sim$ 20 ka ago, to times when Nile perch, crocodiles, and hippos swam in the lake-studded early Holocene Sahara (Kuper and Kröpelin 2006). We will examine the causes and consequences of these Quaternary climatic fluctuations later in this chapter.

Finally, the role of prehistoric and historic human activities cannot be ignored. Trivial during Lower and Middle Palaeolithic times, when prehistoric world population was sparse, there are discernible signs of Upper Palaeolithic impact in many continents, especially through the widespread use of fire. The result was an increase in savanna at the expense of forest, and of grassland at the expense of woodland. With the onset of plant and animal domestication at the start of the Holocene, and the associated rapid increase in world population from roughly 10 million people at the start of the Neolithic (c. 10 ka) to roughly 100 million by 5 ka, increasing exponentially to a billion (10^9) by 2 ka (May 1978), the stage was set for unprecedented deforestation using polished stone axes and adzes initially, later to be replaced with bronze, iron, and steel.

Desertification is the general term used for a change to desert-like conditions in areas previously beyond the climatic limits of existing deserts (Williams and Balling 1996, Williams 2002b). It is brought about by a combination of poor land management (overgrazing, indiscriminate clearing of the vegetation cover, increased soil salinization) and prolonged local or regional intervals of drought. The result is a reduction in plant biomass, animal-carrying capacity, and overall ecological resilience and diversity, with occasionally catastrophic impacts on local pastoral and farming communities Despite eloquent but unsubstantiated claims to the contrary, there is no credible evidence that the deserts depicted on Fig. 28.1 are the result of human action. Sweeping and apparently plausible assertions that the modern deserts are wholly or partly man-made need to be placed in context. The expanding desert syndrome erupts after every severe drought, but the re-establishment of a modern plant cover along the desert margins during the wetter intervals between droughts is seldom mentioned.

Reconstructing Cenozoic Climates in Deserts

Reconstruction of Cenozoic climatic changes in deserts is based on a growing body of evidence from both land and sea, including sediments, plant and animal fossils, isotope geochemistry, landforms, soils, instrumental records, prehistoric archaeology and historical archives. The length of the climatic record attainable from the various independent lines of evidence and the time resolution possible with each of them vary widely, and cover a time range from 10^{-2} to 10^{7} years (Williams et al. 1998). It is therefore critical that the scope and limitations of each type of proxy evidence are thoroughly appreciated. Six very different recent studies drawn from central Asia, North America, the Pyrennes, Tanzania, Indonesia and the Arctic demonstrate the subtleties involved in reconstructing Cenozoic climates and illustrate vividly how our present understanding is still in a state of flux.

Uplift of the Tibetan plateau as a result of the collision of India and Asia $\sim$ 45 Ma ago caused a major change in the distribution of land and sea and was followed by severe desiccation of the region to the north and east of the plateau (the present-day Taklimakan, Badain Jaran and Gobi deserts of China, Inner Mongolia and Mongolia, respectively). The inception or intensification of the Asian winter and summer monsoons has also been attributed to these tectonic events. However, Dupont-Nivet et al. (2007) have developed a fine-resolution magnetostratigraphic chronology for the Eocene-Oligocene transition (33–34 Ma ago) in the Xining basin at the northeastern edge of the Tibetan plateau. Widespread sedimentation in playa lakes persisted during the Eocene and ended abruptly at the Eocene-Ologocene transition, coincident with Cenozoic global cooling at 33–34 Ma associated with the inception of permanent Antarctic ice sheets at that time. The timing of uplift in Tibet is poorly constrained and probably time-transgressive, prompting Dupont-Nivet et al. (2007) to conclude that the desiccation evident at the Eocene-Ologocene transition in this part of Asia is more likely to have been a result of global cooling than of regional tectonic events, although these undoubtedly helped to accentuate aridity.

Zanazzi et al. (2007) arrived at very different conclusions relating to the climatic changes associated with the Eocene-Oligocene transition in the northern Great Plains of central North America. They analysed the stable carbon and oxygen isotopic composition of fossil tooth enamel and fossil bone to obtain a 400 ka record at 40 ka resolution and found that the mean annual temperature fell by $8.2 \pm 3.1°C$ during that time, attributing this drop to an inferred major decrease in the concentration of atmospheric carbon dioxide. They found no evidence of any increase in aridity across the transition, in contrast to the palaeoclimatic inferences of Dupont-Nivet et al. (2007) for central Asia.

The importance for the hydrological cycle of Cenozoic changes in the concentration of atmospheric carbon dioxide is also evident from a recent study from the Pyrenees. A major increase in the concentration of atmospheric carbon dioxide (pCO_2) at the Paleocene-Eocene transition $\sim$55 Ma ago ushered in the warmest 100 ka of the entire Cenozoic. In the Spanish Pyrenees a vast alluvial fan deposited coarse gravels over an area of 500–2000 km^2 within a few thousand years at the start of the Eocene (Schmitz and Pujalte 2007). The gravels imply intense but highly seasonal rainfall at this time, with carbonate nodules within the fan deposits indicating seasonally dry periods. The authors concluded that regional hydrological cycles were highly sensitive to changes in pCO_2.

A further matter for debate in interpreting Cenozoic climates is the issue of long-term Eocene cooling and whether or not it was synchronous from high to low latitudes. Pearson et al. (2007) have re-interpreted previous studies of the oxygen isotopes of deep-sea tropical foraminifera shells using new drill core samples from Tanzania. They were able to demonstrate that in contrast to the exceptionally well-preserved Tanzanian samples, previously analysed deep-sea plankton shells had been diagenetically recrystallised, so that the apparent tropical sea-surface cooling trend inferred by earlier workers was an artefact of diagenetic overprinting. The unaltered Tanzanian foram shells showed a stable warm tropical climate throughout the Eocene, indicating that low latitudes were unresponsive to the long-term Eocene cooling evident in higher latitudes.

The time when continental ice first appeared in Greenland also remains controversial. Eldrett et al. (2007) have recovered macroscopic dropstones from ice-rafted debris from Eocene and Oligocene sediments that were laid down in the Norwegian-Greenland Sea between 38 and 30 Ma ago, with East Greenland as the most likely source. These findings, if correct, show that glaciers were present on Greenland

~20 Ma earlier than previously demonstrated, but at a time when global temperatures and pCO_2 were both relatively high. The timing and the causes of northern high latitude ice accumulation thus remain contoversial.

The late Pliocene increase in aridity evident in East Africa and Ethiopia 3–4 Ma ago (Feakins et al. 2005) has been attributed to closure of the Panama Isthmus and northward diversion of the warm equatorial water which until then had flowed westwards from the Atlantic into the Pacific Ocean. The presence of warm moist air over the North Atlantic coupled to a decrease in insolation linked to increased orbital eccentricity and a decrease in tilt of the earth's axis (leading to cooler high latitude northern summers and milder northern winters) was a prerequisite for widespread and persistent snow accumulation over North America (Williams et al. 1998). The rapid accumulation of ice over North America at 3.5–2.5 Ma was accompanied by global cooling and intertropical aridification.

The causes and consequences of the late Pliocene build up of ice over North America and ensuing tropical climatic desiccation are better constrained than the Paleogene glacial events described earlier but still offer scope for differing interpretations. Cane and Molnar (2001) have proposed that closure of the Indonesian seaway 3–4 Ma ago as a result of northward displacement of New Guinea in the early Pliocene may have triggered a change in the source of water flowing through Indonesia into the Indian Ocean from previously warm South Pacific water to cooler North Pacific water. The concomitant decrease in sea-surface temperatures in the Indian Ocean could have reduced rainfall over East Africa. However, *pace* Cane and Molnar (2001), it seems unlikely that closure of the Indonesian seaway was the sole cause of late Pliocene desiccation in East Africa since the region derives its moisture from both the South Atlantic and the Indian Ocean. We return to the impact of late Neogene cooling and desiccation later in this chapter.

A further *caveat* concerns the processes involved in interpreting past climates. For instance, consider the steps involved in determining the climatic factors responsible for lake level fluctuations. Ideally, we would start with the present-day fluctuations in lake level and relate these to monitored changes in precipitation, evaporation, runoff, inflow, outflow, seepage losses and groundwater inputs. After allowing for the influence of extreme events and their erosional or depositional legacy, the next step would be to use proxy data (e.g., diatoms, trace element geochemistry) to reconstruct past changes in water chemistry (salinity, alkalinity), depth and temperature. A stratigraphically consistent chronology, preferably based on several independent methods (e.g., AMS ^{14}C, OSL, U-series) is essential at each step in the historical reconstruction. Not all methods yield equally precise ages but all must be accurate. To then convert a history of lake level fluctuations to a set of climatic factors is fraught with additional uncertainties. Even assuming that tectonic or non-climatic factors may be ruled out, it is not always possible to decide whether changes in precipitation seasonality outweigh the influence of changes in annual abundance, and whether reduced evaporation offsets any rainfall decrease.

Much past discussion of climatic change in deserts has relied upon geomorphic evidence but great care is needed in using landforms to infer climate, partly because desert landforms are often polygenic and have evolved at different times under ever changing climates. The major erosional landforms in deserts tend to be far older than the depositional landforms. Erosional landforms include desert mountains such as the Hoggar and Tibesti massifs of the Sahara, stony tablelands and plateaux such as the Mesozoic plateaux of central Australia, and denuded lowlands such as the shield deserts of Mauritania and Western Australia, which are located on tectonically stable Archaean and Proterozoic formations with a long history of subaerial erosion.

Depositional landforms include desert dunes, alluvial fans, and floodplains of varying complexity and size, and lacustrine features such as playa lakes. There is often a close relationship between alluvial, lacustrine, and aeolian features, typified by source-bordering dunes emanating from alluvial point-bar sands or by the clay dunes (lunettes) developed on the downwind margins of certain now dry or saline lakes in Algeria, Australia, and Siberia. The distinction between erosional and depositional landforms is often somewhat arbitrary, and will depend upon the temporal and spatial scale in which one is interested. A Holocene diatomite etched by wind erosion is evidence of lacustrine deposition during a less arid interval, as well as evidence of the efficiency of wind erosion during an ensuing drier phase. The focus may be on the deposit (lacustrine diatomite) or on the landform (a series of yardangs eroded from the lake sediments). Alternatively, the emphasis may be on the origin of the lake basin (tectonic or deflationary?) in which the diatomites occur.

Many desert mountains and plateaux are flanked by rectilinear or gently concave rock-cut surfaces or pediments, which are usually mantled with a layer of sediment up to several metres thick. Spectacular stepped flights of pediments are a feature of some of the piedmont slopes of the Atlas, Aurès, Hoggar, and Tibesti mountains in the Sahara and of the Helan Shan in Inner Mongolia. In their distal sectors, the pediments may coalesce and become buried beneath alluvial fans and playa lakes. The larger fans and lake deposits often contain a fragmented but otherwise valuable record of late Tertiary and Quaternary depositional and hydrological fluctuations which may yield useful palaeoclimatic information. Before using the erosional and depositional landforms preserved in our deserts as evidence of past climatic changes, it is essential to distinguish between the effects of climate on the one hand, and those of tectonic and volcanic processes as well as the more subtle controls exercised by lithology and rock structure.

Geomorphic evidence of climatic change is seldom unequivocal and leads too readily to circular argument. Independent verification (or refutation) is vital if palaeoclimatology is to retain credibility. The appropriate question to ask is not whether or not there has been climatic change, but whether or not a particular climate is demanded by a particular suite of landforms. Since most erosional desert landforms can form in a variety of different ways, although the end-product, such as a pediment, may for all practical purposes have the same final appearance, such features will not be usefully diagnostic of past climatic events. More helpful, because more sensitive, is the depositional legacy of former wetter and drier episodes, although here too great caution is needed when invoking climatic change.

Consider, for example, some late Cenozoic lakes in semi-arid Ethiopia. A lake may be created by lava damming a river and may vanish as a result of seismic disruption. Pliocene Lake Gadeb in the semi-arid south-eastern uplands of Ethiopia originated when a lava flow blocked the course of the ancestral Webi Shebele channel some 2.76 m.y. ago (Williams et al. 1979, Eberz et al. 1988). Earlier still, a vast lake occupied what is now the middle Awash valley of the southern Afar Desert and dried out shortly after 3.8–4.0 Ma during an intense phase of tectonic and volcanic activity (Williams et al. 1986). A final and more recent example will suffice to show the climatically ambiguous nature of some lake records. During the height of the prolonged drought which began in 1968 in Ethiopia, at a time when many Ethiopian rift lakes were shrinking, Lake Besaka, a graben lake lying at the foot of Fantale volcano, started to rise at a rate of $0.5\,\mathrm{m\,y^{-1}}$ between 1974 and 1975 (Williams et al. 1981b). A resurgence of hot spring activity along the fault scarps bordering the lake may be the primary cause – possibly a precursor to a future eruption from the caldera, which last erupted about AD 1850. Within two years the brackish lake with its fauna of brine shrimps and flamingoes had become a freshwater lake with a new fauna of freshwater fish and crocodiles.

Only by rigorous multidisciplinary study is it possible to differentiate between desert lakes formerly fed solely by surface runoff (whether local or far-travelled), or solely by groundwater, or by a combination of runoff and groundwater – a combination which will very likely have varied through time (Gasse 1975, Williams et al. 1987, 1991b, Street 1980, Abell and Williams 1989). With these considerations in mind, it is instructive to review the geomorphic history of our two largest tropical deserts: the Sahara and the arid inland of Australia.

Geomorphic History of the Sahara

Precambrian Tectonic Legacy

The geomorphic history of the Sahara begins in the Precambrian, and the influence of the Precambrian tectonic and structural legacy is pervasive across North Africa and the Arabian Peninsula (Black and Girod 1970, Clifford and Gass 1970, Adamson and Williams 1980, 1987, Williams 1984a, Bowen and Jux 1987, El-Gaby and Greiling 1988, F.M. Williams et al. 2004, Thurmond et al. 2004, Guiraud et al. 2005). The Precambrian cover rocks are comparatively undeformed and unaltered, and appear to be of Middle and Upper Proterozoic age. They overlie the weakly to highly metamorphosed and strongly folded and faulted Archaean and Lower Proterozoic formations with a marked erosional unconformity and often form rugged plateaux, mesas, and monoclinal ridges of quartz arenite and conglomerate with intercalated dolerites and basalts. Other parts of Pangaea had a similar history of episodic Archaean and Lower Proterozoic metamorphism followed by prolonged intervals of widespread erosion and late Precambrian sedimentation. For example, the present-day landscapes of the Vindhyan Hills in north central India and the Arnhem Land Plateau in northern Australia differ from some of those in the western Sahara only in now being covered in savanna woodland.

Much of the Precambrian landscape is concealed beneath younger sedimentary and volcanic rocks in places up to 10 km thick, so that the areal extent of outcropping Precambrian formations is confined to about 15% of the Sahara. Despite such limited outcrop, the influence of the ancient basement rocks of the Sahara is out of all proportion to their present surface distribution. In recent decades we have come to appreciate more and more that the patterns of Phanerozoic faulting, rifting, and volcanism, the emplacement of ring-complexes, and the distribution of major depocentres in arid northern Africa and Arabia are a direct reflection of the pervading influence of geological structures that developed during the 650–550 Ma Pan-African orogenic event as well as during the 2000–1250 Ma Eburnean orogeny. Later workers have shown how these early orogenic events have controlled the subsequent location of Phanerozoic uplands and sedimentary basins in the Sahara (Thurmond et al. 2004, Guiraud et al. 2005). Periodic reactivation of some of the major Precambrian structural trends has determined the location of Cenozoic faults, rifts, and volcanoes in northern Africa .

Palaeozoic and Mesozoic Sedimentation

The Pan-African orogeny which reached its climax towards 550 Ma ago was followed by prolonged and widespread erosion which reduced much of the Sahara to a gently undulating plain. Renewed uplift towards 450 Ma heralded the late Ordovician glaciations of the Sahara (Beuf et al. 1971). Around the present Hoggar mountains some of the erosional evidence of these Ordovician ice caps, such as the glacially striated pavements, is so well preserved that it almost appears to be Pleistocene. Melting of the Saharan ice caps was followed by the rapid and world-wide early Silurian rise in sea level, comparable to the glacio-eustatic sea-level rise which followed melting of the late Pleistocene ice caps. Accumulation of Carboniferous mudstones, limestones and sandstones across the Sahara eventually gave way to Hercynian uplift in the western Sahara and Atlas associated with opening of the North Atlantic. Uplift and folding of the previously horizontal Palaeozoic sedimentary rocks triggered the Triassic and later erosion during which the woodlands of the northern Sahara were rapidly buried beneath the sandstones and mudstones laid down by Mesozoic rivers flowing from the south and east. Silicified tree-trunks, some still in upright positions of growth, are vivid indications of the speed and effectiveness of the Mesozoic fluvial aggradation. A hundred million years later, the prehistoric peoples of the Pleistocene Sahara were to use fragments of this silicified wood for stone toolmaking.

The early Mesozoic opening of the South Atlantic and the ensuing separation of the African and South American lithospheric plates resulted in a series of Mesozoic marine transgressions. Much of the Sahara was relatively flat and low-lying at that time, and flanked by warm and shallow epi-continental seas. One such marine incursion flooded much of the central Sahara during the Upper Cenomanian and Lower Turonian, although the Hoggar uplands remained above sea level, as did much of the West African craton.

These Mesozoic marine and non-marine sedimentary formations comprise some of the greatest present-day aquifers in North Africa, and include the Nubian Sandstone Formation of the eastern Sahara and the 'Continental intercalaire' (Kilian 1931) of the central Sahara. The more or less horizontal Mesozoic formations are scattered across the Sahara today in the form of extensive sandstone and limestone plateaux or hamada. Australia had a similar Mesozoic history, and there is little to distinguish the great stony tablelands and associated gibber plains of the present Australian desert from the vast gravel-strewn hamada and stony reg surfaces of the Sahara.

A classic example of a Saharan sandstone plateau or hamada is the Gilf Kebir in the now hyperarid desert of south-eastern Libya and south-western Egypt. The margins of the Gilf Kebir as well as those of other great sandstone plateaux in southern Libya are highly crenulated and deeply dissected by now inactive river valleys. Peel (1966) and many later observers have emphasized the former efficiency of such fluvial erosion (Griffin 2006), but perhaps the most eloquent testimony to these pluvial episodes – apart from the dry valleys themselves – are the great galleries of Upper Palaeolithic and Neolithic rock paintings and engravings of now vanished herds of elephants, giraffes, and domesticated cattle (Muzzolini 1995, Coulson and Campbell 2001). More recently, Breed and her colleagues have made brilliant use of shuttle-imaging radar to identify Pleistocene and older river channels in the eastern Sahara (Breed et al. 1987). These now defunct watercourses flowed at least intermittently during the early to middle Pleistocene and at intervals thereafter, including the early to middle Holocene. During the early and middle Pleistocene, small bands of *Homo erectus* roamed the Sahara equipped with their all-purpose Acheulian toolkit of bifacially worked hand axes, cleavers and scrapers (Breed et al. 1987, McHugh et al. 1988). By Holocene times, the Palaeolithic hunters and gatherers had been replaced by Neolithic pastoralists who grazed their cattle throughout the Sahara (Williams and Faure 1980, Clark and Brandt 1984, McHugh et al. 1989).

Cenozoic Deep Weathering, Uplift, and Erosion

Withdrawal of the shallow, equatorial Cretaceous seas from the Sahara was followed by a very long interval of intense early Tertiary weathering and leaching of the forested lowlands of the tectonically quiescent southern Sahara (Faure 1962). Near-surface solution and redeposition of iron and silica during the Palaeocene and Eocene gave rise to the resistant caprocks of ironstone or silicified rock which now protect many of the Mesozoic and Tertiary plateau summits from erosion. The present-day geomorphic outcome is a process of slow and episodic scarp retreat by undercutting of the less resistant mudstones and softer sandstones during wetter phases, followed by collapse of the resistant caprocks. The undercutting is effective even today, and is aided by seepage at the cliff base, by salt weathering and by chemical weathering and deflation. Similar processes of scarp retreat have been invoked to explain boulder-mantled debris slopes in the semi-arid northwest of South Australia.

Post-Eocene uplift triggered a widespread phase of mid-Tertiary erosion within major massifs such as Tibesti and the Hoggar, as well as in more isolated ring-complexes such as Jebel Arkenu and Jebel ‘Uweinat in south-east Libya or Adrar Bous in central Niger. The mid-Tertiary drainage system appears to have been a highly efficient and well-integrated system which kept pace with the various epeirogenic uplifts across the Sahara. The Nile cut down through Nubian Sandstone capping the Sabaloka ring complex to form the Sabaloka gorge north of Khartoum – one of the many instances of superimposed Cenozoic drainage in the Sahara (Grove 1980, Williams and Williams 1980, Thurmond et al. 2004). The early Tertiary mantle of deeply weathered rock was virtually removed from the uplands of the southern Sahara, leaving a bare and rugged landscape of gaunt rocky pinnacles and boulder-mantled slopes. Episodic deep weathering followed by episodic erosion and exhumation of the weathering front became the geomorphic norm of the later Cenozoic (Dresch 1959, Thorp 1969, Williams 1971). There seems little doubt that mid-Tertiary tectonic movements performed a dominant role in the initial pulse of erosion but the late Cenozoic climatic oscillations became increasingly important erosional pacemakers thereafter (Williams et al. 1987).

The sandy colluvial–alluvial debris eroded from the Saharan uplands was carried away from the mountains by the Tertiary and early Quaternary rivers to be in part deposited in late-Cenozoic marine deltas such as those of the Nile, the Niger, and the Senegal. However, a considerable proportion of the sediment began to accumulate in the closed interior basins created during the course of late Mesozoic and Cenozoic faulting, rifting and epeirogenic movements.

It was the unconsolidated Tertiary sediments laid down in large subsiding sedimentary basins such as the Kufra-Sirte basin in Libya, or the Chad basin, which provided the source material for the late Tertiary and Quaternary desert dunes. Miocene tectonic uplift in East Africa may have contributed to the desiccation in this region from about 8 Ma onwards (Sepulchre et al. 2006). In the Chad basin, Servant (1973) identified wind-blown sands in a number of very late Tertiary stratigraphic sections. He concluded that the onset of aridity and the first appearance of desert dunes in this part of the southern Sahara was a late Tertiary phenomenon. Using fossil and sedimentary evidence, Schuster et al. (2006) have since confirmed that the onset of recurrent desert conditions in the Chad basin began at least 7 Ma ago. Further north, in the Hoggar, some elements of the late Tertiary flora were already physiologically well adapted to aridity (Maley 1980a, b, 1996). If we accept the sedimentological evidence of Servant (1973) and of Schuster et al. (2006) and the palynological evidence of Maley (1996), then it follows that the onset of climatic desiccation and the ensuing disruption of the integrated mid-Tertiary Saharan drainage network (Griffin 2006) was a feature of the very late Tertiary, long pre-dating the arrival of *Homo sapiens*. Before pursuing this topic in greater detail, we need to retrace our steps and examine the impact of Cenozoic volcanism, faulting, and rifting upon the Saharan landscape.

Cenozoic Volcanism, Uplift, and Rifting

The basin and swell topography of the present-day Sahara (Fig. 28.2) is very largely a function of the Tertiary epeirogenic movements discussed earlier. An additional and extremely important factor was the massive extrusion of lava which accompanied and accentuated the late Cenozoic uplift of all the existing uplands, including the Ethopian Highlands, Jebel Marra, Tibesti, the Hoggar, and the Aïr.

An estimated 8000 km^3 of volcanic rock was erupted during the formation of Jebel Marra, and about half that amount (3000 km^3) during the eruptions which gave rise to Tibesti (Francis et al. 1973). The volcanic eruptions which helped to create Jebel Marra (elevation: 3088 m), Tibesti (3415 m) and the Hoggar (2918 m) were associated with significant uplift of the basement rocks which now lie beneath the late Cenozoic volcanic rocks (Guiraud et al. 2005). How much of this uplift took place during as opposed to before the eruptions is not known, but of the uplift itself there seems no doubt. Relative to the basement rocks on the adjacent plains, the basement rocks of Tibesti and the Hoggar have risen about 1000 m and those of Jebel Marra about 500 m (Bordet 1952, Vincent 1963, Vail 1972a,b). Much of the volcanic activity is of Miocene age and younger, and significant erosion of the crystalline basement rocks immediately preceded the volcanic activity and its erosional aftermath (Rognon 1967, Williams et al. 1980).

The timing of uplift in the Aïr massif is less well established (Williams et al. 1987), but the presence of possible Cretaceous sedimentary rocks at 1400 m on Tamgak granite plateau and very high (c. 1500 m) on the Goundai ring-complex (Raulais 1951, Vogt and Black 1963) suggests a possible 1000–2000 m of post-Cretaceous uplift. Given the late Cenozoic updoming of the Hoggar (Bordet 1952, Rognon 1967) and Jebel Marra (Vail 1972a,b) it seems likely that uplift of the Aïr is also of late Cenozoic age. The Mesozoic sedimentary formations south and west of Agadès all dip westwards away from the main Aïr massif and are again consistent with post-Cretaceous updoming of the massif.

Uplift of the Ethiopian Highlands is also a late Cenozoic event (Pik et al. 2003). Pollen grains characteristic of tropical lowland rainforests are abundant in Miocene lignites intercalated between basalt flows dated by K/Ar content at 8 Ma (Yemane et al. 1985). The basalts crop out at roughly 2000 m elevation near Gondar in the north-western Ethiopian uplands, suggesting roughly 2000 m of uplift in the past 8 m.y., or an average of 0.25 mm y^{-1}. The great depth and steepness of the Blue Nile gorge and the Tekazze gorge in central western Ethiopia is consistent with late Cenozoic uplift, and Faure has argued for an acceleration of uplift from Tertiary to Quaternary based on uplift rates of coral reefs along the Red Sea (Faure 1975). Since there is no necessary link between coastal uplift and uplift of the Ethiopian Highlands, the question is best left open. One thing is certain, however, and that is the enormous volume of rock eroded and removed from Ethiopia since extrusion of the Upper Oligocene and Miocene Trap Series basalts. The Blue Nile, the Tekazze, and their tributaries have removed about 100 000–200 000 km^3 of rock from the north-western Ethiopian uplands in the last 10–20 m.y., which is also equivalent to the volume of the Nile cone in the eastern Mediterranean (McDougall et al. 1975, Williams and Williams 1980). A comparison of modern rates of erosion in upland Ethiopia and mean geological rates of denudation persuaded Williams and Williams (1980) that uplift of the Ethiopian Highlands was episodic, with prolonged intervals of relative stability between shorter intervals of rapid uplift. The abundance of montane forest podocarpus and juniper pollen in Pliocene sediments, which are now at low elevations in the lower Awash valley of the west-central Afar Desert (Bonnefille et al. 2004) also indicates that uplift of the Ethiopian uplands was accompanied by down-faulting

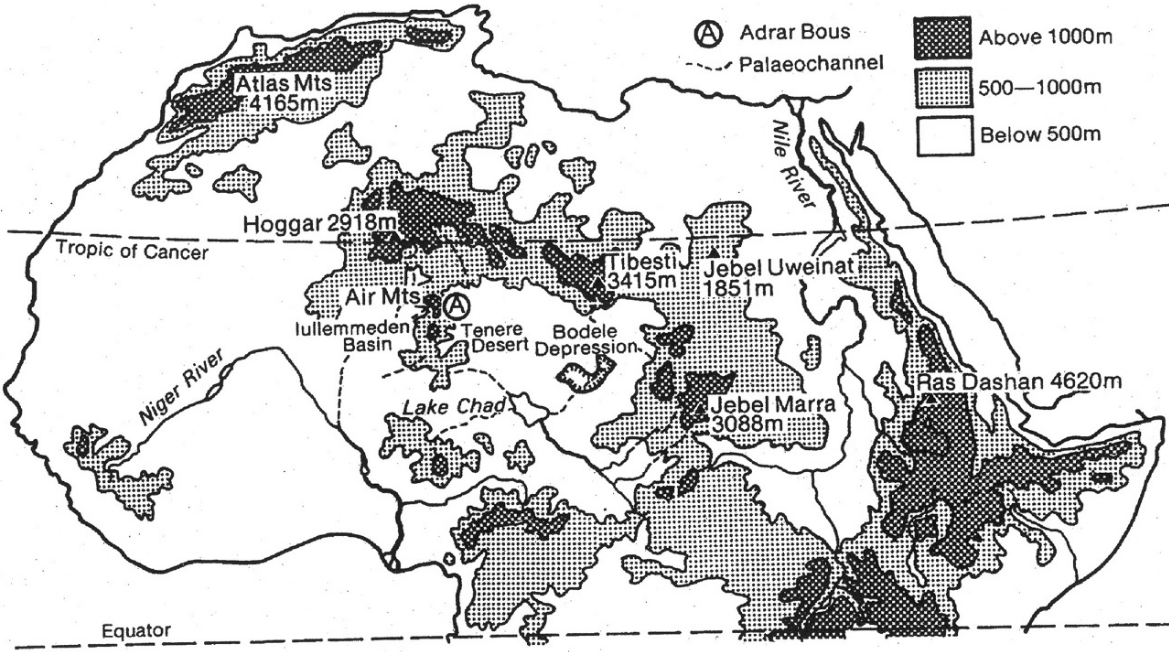

Fig. 28.2 Relief map of northern Africa, showing distribution of major Saharan uplands and lowlands. Also shown is the locatioan of Adrar Bous ring-complex in the heart of the Sahara, immediately east of the northern Aïr massif, in the Tenere Desert of Niger (after Williams et al. 1987, Fig. 1)

and subsidence along the adjacent margins of the eastern Ethiopian escarpment (Adamson and Williams 1987).

The pattern of rifting in Ethiopia has created a very particular type of desert in the Afar Depression (Tiercelin 1987, Williams et al. 1986, Adamson and Williams 1987). The Afar is a lava desert and is one of the hottest and most forbidding deserts on Earth. Flanked by the mighty Ethiopian escarpment on the west, it descends over 150 m below sea level at Lake Asal, a salt lake now separated from the adjacent Red Sea by a puny Pleistocene lava dam.

Neogene fluvio-lacustrine sediments in southern Afar have yielded one of the longest, richest, and most complete records of fossil hominid evolution anywhere in Africa (Clark et al. 1984). Associated with these australopithecine hominid fossils are abundant superbly preserved Pliocene fossils of elephants, pigs, bovids, hippos, crocodiles, and sundry non-aquatic carnivores. The fauna is a savanna fauna, and most taxa belong to now extinct species. The presence of such fossils in parts of the Afar that do not now support much life, together with the presence of thick lacustrine deposits within the Neogene formations again raises the issue of possible late-Cenozoic climatic desiccation, a topic to which we now turn.

Late Cenozoic Desiccation

Tertiary volcanism in the central and southern Sahara was preceded and accompanied by prolonged deep weathering. In central Niger kaolinitic and bauxitic weathering profiles up to 45 m thick are developed on rocks of Eocene to Precambrian age. Uplift in the mid-Tertiary resulted in a change from previously biogenic and chemical sedimentation in this region to dominantly clastic sedimentation (Faure 1962, Greigert and Pougnet 1967, p. 157). Rejuvenated rivers flowing down from the great watersheds of Tibesti, the Hoggar, and the Aïr deposited the fluvial gravels, sands, and clays of the 'Continental terminal' extensively around their parent uplands. Williams and co-workers have concluded that in Niger and adjacent areas 'the origin of the Sahara as a continental desert. . . may be said to stem from the Miocene Alpine orogeny and the subsequent stripping of the Eocene deep weathering profile' (Williams et al. 1987, p. 109). Apart from tectonic uplift, what other factors were responsible for this dramatic change from a landscape of lowland equatorial rainforest to one of bare rocky inselbergs and desert dunes?

A major influence, not mentioned so far, was the post-Palaeozoic northward drift of the African lithospheric plate. Triassic Africa was part of the Gondwana supercontinent, as were South America, Antarctica, Australia, and India. With the Jurassic and earlier separation of Gondwana into the two continents of West Gondwana (Africa and South America) and East Gondwana (Australia, Antarctica, and India) the stage was set for further break-up of these two large continents during the Cretaceous (Owen 1983).

The Cretaceous equator in Africa ran from southern Nigeria through central Chad and the northern Sudan into Arabia. During the late Mesozoic and Cenozoic, the African plate moved northward with a slow clockwise rotation (Habicht 1979,Owen 1983, F.M. Williams et al. 2004). Early Cenozoic Africa was south of its present position by only a few degrees of latitude and came into contact with Europe as a result of a slight clockwise rotation during the Miocene and Pliocene. One outcome of the ensuing crustal deformation was the uplift of the Atlas, noted earlier, which was also coeval with volcanism and updoming of the Hoggar, Tibesti, Aïr and Jebel Marra uplands, creating the major elements of the topography depicted on Fig. 28.2.

A further outcome of Africa's Mesozoic and Cenozoic northward drift and rotational movement was a corresponding southward shift of the equatorial rainforest. This zone once ran obliquely across the Sahara from Egypt and the northern Sudan south-westwards towards southern Nigeria. Aridity set in earlier in Morocco, Algeria, and Tunisia than in Egypt and the Sudan, as is evident from the abundance of Mesozoic and younger evaporite formations in the north-western Sahara, which by then had already reached dry tropical latitudes (Coque 1962, Conrad 1969, Williams 1984a).

Three additional influences contributed to the late Cenozoic desiccation of the Sahara. These were the uplift of the Tibetan plateau, the build-up of continental ice in Antarctica and the Northern Hemisphere, and cooling of the world's oceans. We consider the possible causes of these phenomena in the final section of this chapter; our concern here is purely with their effects upon the Sahara.

Late Cenozoic uplift of the Himalayas and of the vast Tibetan plateau was associated with the intensification (and perhaps the inception) of the easterly jet stream which today brings dry subsiding air to the deserts of Arabia and northern Africa. A major change in the flora and fauna of the Potwar plateau in the Siwalik foothills of Pakistan between 7.3 and 7.0 Ma may also be related to Himalayan uplift and is consistent with intensification of the Indian summer monsoon, if not with its origin at that time (Quade et al. 1989, Cerling et al. 1997).

Accumulation of continental ice in Antarctica may seem somewhat remote from Saharan desiccation but was in fact of critical importance. Mountain glaciers were present on Antarctica early in the Oligocene, and a large ice cap was well established by 10 Ma (Shackleton and Kennett 1975). Continental ice was slower to form in the Northern Hemisphere but was present in high northern latitudes by 3 Ma, and possibly by 5 Ma or even well before then, with a rapid increase in the rate of ice accumulation towards 2.5 to 2.4 Ma (Shackleton and Opdyke 1977, Shackleton et al. 1984). As temperatures declined over the poles, and sea surface temperatures at high latitudes grew colder, the temperature and pressure gradients between the Equator and the poles increased. There was a corresponding increase in Trade Wind velocities, and hence in the ability of these winds to mobilize and transport the alluvial sands of the Saharan depocentres and to fashion them into desert dunes. Higher wind velocities were also a feature of glacial maxima during the Pleistocene and were responsible for transporting Saharan desert dust far across the Atlantic (Parkin and Shackleton 1973, Parkin 1974, Williams 1975, Sarnthein 1978, Sarnthein et al. 1981). During the Last Glacial Maximum centred on 21 ± 2 ka (Mix et al. 2001) Australian desert dust was also blown as far as central Antarctica (Petit et al. 1981).

Late Cenozoic cooling of the ocean surface was also responsible for reducing intertropical precipitation. Galloway (1965) noted that two-thirds of global precipitation now falls between latitudes 40°N and 40°S and depends upon effective evaporation from the warm tropical seas. The ocean surface cooling, which was linked to global cooling associated with high-latitude continental ice build up and enhanced cold bottom-water circulation, would help to reduce evaporation from the tropical seas, thereby reducing rainfall across North Africa.

The late Cenozoic desiccation which created the largest desert in the world was therefore a result of a number of factors. Northward drift of the African plate ultimately helped to disrupt the warm Tethys Sea with its abundant supply of moist maritime air. Northern Africa moved away from wet equatorial latitudes into the dry subtropics. Growth of the great continental ice sheets and cooling of the oceans saw a decrease in precipitation and an increase in the strength of the Trade Winds. At the start of the Oligocene there was a sharp drop in sea surface temperatures, with eventual global repercussions.

Sudano-Guinean woodland covered much of the Sahara during the Oligocene and early Miocene, having replaced the equatorial rainforest of Palaeocene and Eocene times. During the late Miocene and early Pliocene a xeric flora, well adapted to aridity, began to replace the earlier woodland, so that many elements of the present Saharan flora were already present during the late Pliocene, when aridity became even more severe across the Sahara and the Horn of Africa (Bonnefille 1976, 1980, 1983, Maley 1980b).

The combination of a reduction in plant cover and a trend towards more erratic rainfall had a profound impact on the late Cenozoic rivers of the Sahara (Griffin 2006). Big rivers capable of carving large valleys became seasonal or ephemeral. Integrated drainage systems became segmented and disorganized. Wind mobilized the sandy alluvium into active dunefields. Dunes formed barriers across river channels no longer competent to remove them. Dust storms left the desert topsoils depleted in clay, silt, and organic matter. The Sahara was now a true wilderness, as the Arabic word implies.

Quaternary Climatic Fluctuations

Mid-Pliocene closure of the Panama isthmus towards 3.2 Ma paved the way for the rapid accumulation of continental ice sheets in high northern latitudes during the late Pliocene (Schnitker 1980, Loubere and Moss 1986, Prentice and Denton 1988). Oxygen isotope evidence from deep-sea cores indicates that the onset of major Northern Hemisphere continental glaciations at 2.4 ± 0.1 Ma (Shackleton et al. 1984) also coincided with cooling in high southern latitudes (Kennett and Hodell 1986). The 2.3–2.5 Ma temperature drop is also evident in the south-eastern uplands of Ethiopia (Bonnefille 1983) and the dry northern interior of China, with the beginning of widespread loess accumulation in the Loess Plateau of central China dated to 2.4 Ma (Heller and Liu 1982). In the north-western Mediterranean region, the presence of a Mediterranean vegetation adapted to winter rains and summer drought is already evident at 3.2 Ma, but it is not developed in its modern form until about 2.3 Ma (Suc 1984).

Magnetic susceptibility measurements of deep-sea cores from the Arabian Sea and the eastern tropical Atlantic also reveal a change in the length of astronomically controlled climatic cycles at this time. Before 2.4 Ma, the dominant cycles are the 23-ka and 19-ka precession cycles, but after 2.4 Ma, the 41-ka obliquity cycle becomes dominant (Bloemendal and deMenocal 1989). Although the boundary between Pliocene and Pleistocene is now defined by the International Union of Geological Sciences as 1.8 Ma, this somewhat arbitrary date should not obscure the fact that the continental glaciations characteristic of the Quaternary were ushered in by the dramatic fall in global temperatures towards 2.4 Ma.

There is now widespread recognition that the magnitude and frequency of the late Pliocene and Quaternary glaciations were strongly influenced by orbital perturbations. This recognition is thanks to the work of the brilliant Yugoslav astronomer and mathematician, Milutin Milankovitch, who was the first to persuade Quaternary scientists of the climatic importance of the Earth's orbital variations (Milankovitch 1920, 1930, Chappell 1974, Imbrie and Imbrie 1979, Berger 1981, Williams et al. 1998). The major cycles identified by Milankovitch, and for which he calculated the changes in insolation received on Earth at different seasons and latitudes for the successive stages of each cycle, are as follows. The 100-ka orbital eccentricity cycle is determined by the changing elliptical path of the Earth around the Sun. The changing tilt of the Earth's rotational axis gives the 41-ka obliquity cycle. The precession of the equinoxes varies with the changing distance between Earth and Sun and gives the 23-ka cycle.

Although the correlations between orbital perturbations, ice volume fluctuations, and glacio-eustatic sea level fluctuations are statistically significant and now well accepted (Hays et al. 1976), the relative influence of the various cycles has varied during the course of Quaternary time.

Scrutiny of oxygen isotope records from deep-sea cores spanning the full duration of the Quaternary persuaded Williams et al. (1981a) that the early Pleistocene from 1.8 to 0.9 Ma was subject to high-frequency but low-amplitude fluctuations in the oxygen isotope differences between glacial and interglacial maxima. In contrast, the last 0.9 Ma were characterized by high-amplitude but low-frequency fluctuations in oxygen isotopic composition (Williams et al. 1981a). Since changes in the oxygen isotopic composition of benthic Foraminifera are very broadly a reflection of changes in global ice volume (Shackleton 1977, 1987), the changing pattern of glaciation during the Quaternary will also be reflected in variations in the severity and frequency of cycles of glacial aridity in the tropics.

Ruddiman and Raymo (1988) demonstrated that the 100-ka orbital eccentricity cycle was dominant in North Atlantic cores during the last 0.78 Ma (i.e., during the Brunhes magnetic chron). Before then, during the Matuyama chron from 2.60 to 0.78 Ma, the 41-ka orbital cycle was dominant, reflected in more frequent but lower amplitude climatic fluctuations.

Given the powerful influence exerted by the North Atlantic upon both ice volume and precipitation in the Northern Hemisphere, certain palaeoclimatic inferences may be drawn with respect to the Pleistocene Sahara. We have long known that the last glacial maximum in the Sahara was a time of accentuated aridity, with reactivation (or advance) of desert dunes up to 500–1000 km beyond their present southern limits (Grove 1958, Grove and Warren 1968, Talbot 1980, Williams 1975, 1985). During these times of glacial aridity and desert expansion, vast plumes of Saharan and Arabian desert dust were mobilized and blown far out to sea. Over the past 0.6 m.y., maximum concentrations of Saharan desert dust in equatorial Atlantic deep-sea cores coincide with times of low sea-surface temperature or glacial maxima (Parmenter and Folger 1974, Bowles 1975). A similar pattern of glacial aridity is evident in the Red Sea and Gulf of Aden. Planktonic Foraminifera from deep-sea cores collected in this region reveal through their changing isotopic composition that during the last 250 000 years, at least, glacial maxima were times of extreme aridity, with much increased sea-surface salinity reflecting even higher local rates of evaporation than today (Deuser et al. 1976).

It would be misleading to portray all arid phases as coinciding with glacial maxima and all humid phases with peak interglacial times. The reality is more complex. Lake levels in Lake Chad (Servant 1973) and Lake Abhe (Gasse 1975) were high for at least 20 000 years before 18 ka when they fell rapidly. Lake Abhe remained dry until 12 ka, and Lake Chad intermittently dry until then, after which they both rose again rapidly, reaching peak levels towards 9 ka. Since about 4.5 ka these lakes have remained relatively low, with occasional short-lived transgressions. Very schematically, we could consider the interval of high lake levels from 30 to 18 ka as representing a humid glacial phase, the interval of low lake levels from 18 to 12 ka as an arid glacial phase; the interval of early Holocene high lake levels as a humid interglacial phase; and the interval of late Holocene relatively low lake levels as a dry interglacial phase. Even this fourfold subdivision is a caricature of reality, and does not take into account local hydrological and geomorphic controls over precipitation, runoff, evaporation, and groundwater inflow and seepage (Fontes et al. 1985, Abell and Williams 1989).

Whatever their ultimate causes (Kutzbach and Street-Perrott 1985, Gasse et al. 1990, Street-Perrott and Perrott 1990), the consequences of the alternating wetter and drier Quaternary climatic phases are very evident throughout the Sahara. For instance, at the isolated ring-complex of Adrar Bous in the geographical heart of the Sahara (Fig. 28.2), the geomorphic expressions of these past climatic fluctuations include active and stable dunes, lake strandlines and partially deflated lacustrine diatomites, alluvial fans, alluvial terraces, and partly buried palaeochannels and former backswamps (Fig. 28.3). The stratigraphic evidence is equally informative and extends well back into the middle Pleistocene (Williams et al. 1987). Phases of rapid erosion with associated deposition of coarse sands and gravels alternated with longer intervals of minimal erosion, fine-grained sedimentation in low-energy environments, and soil development in and around the mountain. The presence at Adrar Bous of late Pleistocene and Holocene freshwater snails and gastropods allows us to use palaeoecological and isotopic evidence (Williams et al. 1987) to test the inferences drawn from sedimentology and geomorphology, and additional evidence is yielded by prehistoric stone tool assemblages, hearths, graves,

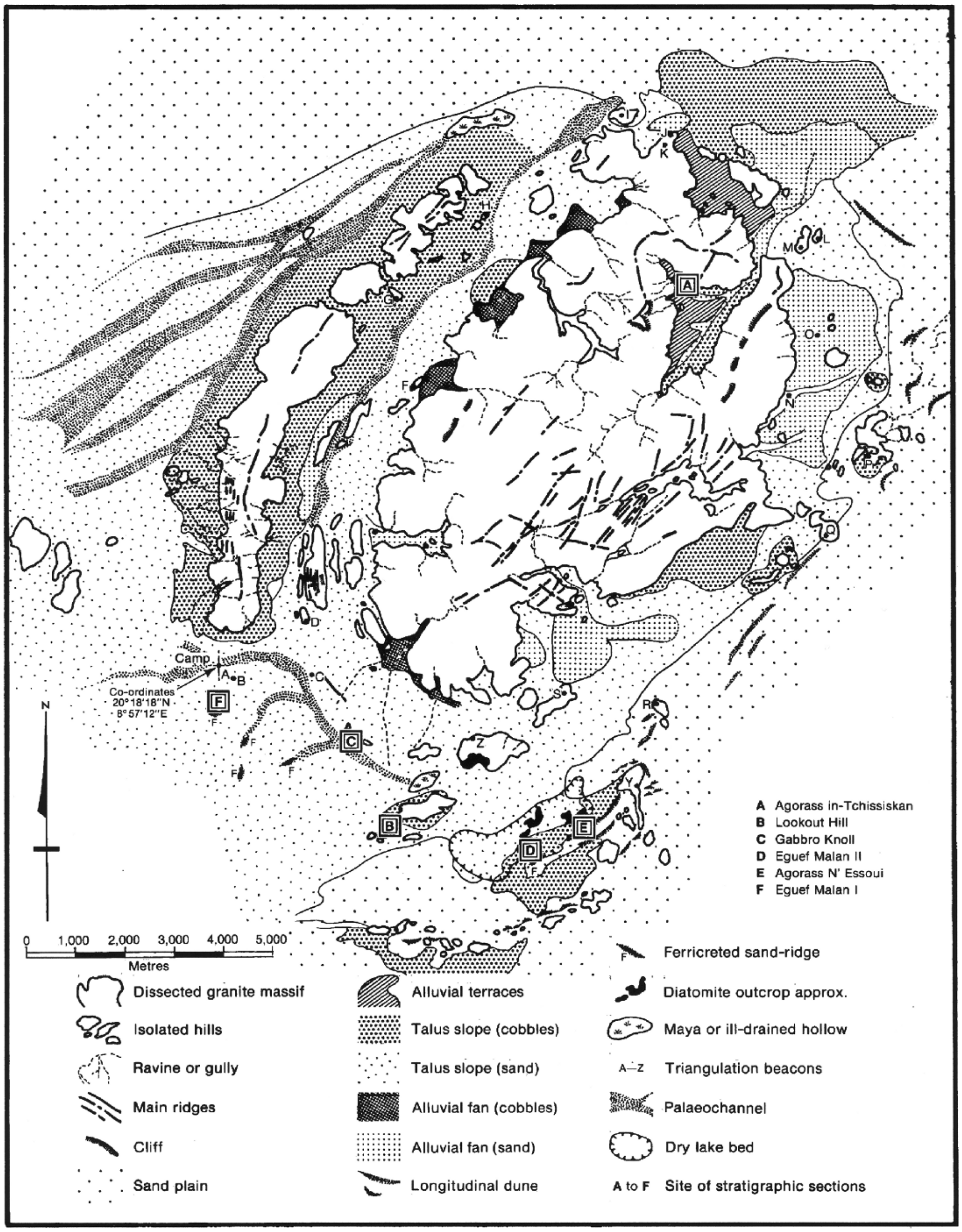

Fig. 28.3 Geomorphic map of Adrar Bous ring-complex in the Tenere Desert of Niger, showing the major landforms and location of the stratigraphic sections discussed in the text (after Williams et al. 1987, Fig. 3)

and middens, the last with burnt remains of locally consumed fauna (Clark et al. 1973, Smith 1980).

Similar Quaternary interdisciplinary or multidisciplinary studies are now the norm in different parts of the Sahara, so that despite its vastness and periodic difficulties of access, the Sahara has provided the best dated and most comprehensive evidence of late Quaternary environmental fluctuations so far available from any of our deserts (Maley 2000, Hoelzmann et al. 2004).

Geomorphic History of the Australian Desert

The following account is distilled from Bowler (1973, 1976, 1998), Shackleton and Kennett (1975), Williams (1984b,c, 1991, 2000, 2001), Chen (1989), Williams et al. (1991a, 2001), Alley et al. (1995), Ayliffe et al. (1998), Bowler and Price (1998), Magee (1998), Bowler et al. (1998, 2001), Veevers (2000), McGowran et al. (2004), Fujioka et al. (2005), Martin (2006), Prideaux et al. (2007), Revel-Rolland et al. (2006), van der Kaars et al. (2006) and Sniderman et al. (2007).

Precambrian Tectonic Legacy

Until about 2500 Ma, much of what is now western and northern Australia consisted of a granitic basement of Archaean rocks. Localized faulting and rifting created a vast regional sediment trap in northern Australia towards 2500 Ma. Over the next few hundred million years, intercalated marine and non-marine gravels, sands, and muds, together with volcanic rocks, accumulated within this depocentre, attaining a thickness of over 14 000 m. Major regional metamorphism ensued towards 1870–1800 Ma, during which the Lower Proterozoic sedimentary and volcanic formations, together with some of the adjacent basement rocks, were folded, faulted, and metamorphosed to form the Lower Proterozoic metasediments which crop out today as rocky strike-ridges in northern Australia. This period of regional metamorphism was followed by uplift, erosion, and by several minor episodes of granite intrusion and volcanic activity. For several tens of millions of years thereafter, prolonged erosion, interrupted by episodic faulting and uplift, ultimately created a Precambrian landscape of gently undulating relief with sporadic hills and shallow valleys. Thereafter, the region has been tectonically stable apart from slow epeirogenic uplift during the late Phanerozoic.

Several hundred metres of horizontal sands, interbedded lavas, and minor basal gravels were laid down across the early Precambrian land surface towards 1690–1650 Ma. These Middle and Upper Proterozoic sandstones are the older Precambrian cover rocks in northern Australia and today form rugged sandstone plateaux such as the Arnhem Land plateau with its joint-controlled gorges and steep erosional cliffs. These Precambrian cover rocks or plateau sandstones have protected the underlying Lower Proterozoic metasediments from subsequent erosion. Slow scarp retreat has gradually exhumed the original Precambrian topography, so that the present relief of bevelled strike ridges and undulating rock-cut surfaces is, in parts, a resurrected and, except for the modern vegetation cover, a virtually unmodified Lower Proterozoic landscape.

Palaeozoic and Mesozoic Sedimentation

Palaeozoic sandstones and limestones crop out in the less elevated parts of northern Australia but are conspicuously absent from the stable cratonic areas of western Australia as well as from the summits of the northern Precambrian plateaux.

During the early Cretaceous, towards 135–100 Ma, Australia consisted of three large islands separated by shallow seas. The two western islands were the Precambrian shield region of western Australia and the Precambrian and Palaeozoic uplands of northern and central Australia, indicating that these upland areas have had a long history of subaerial erosion.

Some of the present northern Precambrian plateaux, such as the Arnhem Land plateau, have a thin cover of Cretaceous sediments. These Cretaceous mudstones and sandstones also mantle the Lower Proterozoic and Archaean rocks north and west of Arnhem Land, where they are up to 200 m thick. Scarp retreat of the plateau has probably been accelerated by the uplift which followed the 110 Ma Aptian marine transgression, and may be a Cenozoic rather than a Cretaceous phenomenon, since the Cretaceous sediments are uniformly and deeply weathered wherever they occur. In central Australia, the horizontal Mesozoic formations often have a caprock of silcrete. Further north, in what are now the seasonally wet tropics, the caprock consists of ferruginous ironstone (laterite or ferricrete) or even of bauxite. Such deep weathering implies prolonged and efficient leaching under conditions of slow or ineffective mechanical erosion, and normally requires low gradients, high effective precipitation and a dense vegetation cover.

Cenozoic Deep Weathering, Uplift, and Erosion

By late Cretaceous time, the sea had retreated from most of Australia, and in the centre and north there were several prolonged intervals of deep weathering of the newly emerged land surface. This prolonged weathering under conditions of tectonic stability was interrupted by minor intervals of mostly gentle uplift with local faulting and folding, notably in the middle to late Miocene.

At least three post-early Cretaceous erosion surfaces have been identified by different workers in central and northern Australia on the basis of elevation, laterite-capped remnants, and regional slope. Assuming that block-faulting can be ruled out, and assuming that the weathering mantle did indeed develop on a gently sloping surface, both of which at present are unproven assumptions, then the presence of such stepped surfaces is entirely consistent with episodic Cenozoic uplift and associated vertical and lateral erosion. The geomorphic evidence accords well with such an interpretation. Rejuvenation of the drainage in northern Australia has led to vertical incision and the development of rugged relief consisting of steep valleys, rocky strike-ridges, boulder-mantled granite hills, and steep-sided sandstone plateaux.

Late Cenozoic Desiccation

Until about 130 Ma, Australia, Antarctica, and Greater India were all part of East Gondwana. Very early in the Cretaceous a rift developed between Greater India and Australia–Antarctica, and the present Indian Ocean began to form. Sea-floor spreading was very rapid between 80 and 53 Ma, attaining rates up to 175 mm y^{-1}. The initial separation of Australia from Antarctica at 90 Ma was at first very slow and remained so until about 30 Ma. During the last 30 m.y., Australia has drifted north at a mean rate of roughly 50–70 mm y^{-1}. Concomitant rifting along the southern continental margin and the eastern edge of continental Australia led to separation of the Campbell Plateau, New Zealand, and the Lord Howe rise from Australia, and formation of the Tasman Sea, and, ultimately, of the Southern Ocean.

As Australia drifted north, the Eastern Highlands were the centre of sustained volcanism, although the locus of volcanic activity shifted south as Australia moved north. The Eastern Highlands were uplands well before the Tertiary, and sporadic uplift continued during the late Cretaceous and Cenozoic, even if the detailed pattern of uplift is still unclear. The west-flowing drainage of eastern Australia is thus very old, and great river basins such as the ancestral Murray–Darling were already in existence during the late Cretaceous.

In western Australia, early Cretaceous rifting, volcanism, uplift and later subsidence accompanied the separation of Greater India from Australia. Late Palaeocene to Eocene rifting also accompanied the separation of Australia from Antarctica. As a result of these tectonic upheavals, the once integrated network of drainage in western Australia became progressively disrupted from 130 to 40 Ma (early Cretaceous to late Eocene), and the rivers became less and less active during the Oligocene and early Miocene. By mid-Miocene times (15–10 Ma), a conspicuous network of linear salt lakes occupied the western half of the Australian continent, a witness to early Cenozoic tectonic disruption and later Cenozoic climatic desiccation.

As Australia moved north away from Antarctica, it came increasingly under the influence of tropical climatic systems and dry, subsiding, anticyclonic air masses. The long-term reduction in precipitation and runoff is reflected in the late Cenozoic trend towards the endoreic and areic drainage systems which are today characteristic of roughly two-thirds of the continent. Cooling of the Southern Ocean accentuated the trend towards aridity evident in the changing flora and fauna of inland Australia. Figure 28.4 shows a fluctuating decline in Southern Ocean sea-surface temperatures from roughly 19°C in the early Eocene to only 7°C in the Oligocene, associated with the progressive accumulation of ice in Antarctica discussed earlier in this chapter. The global cooling and ensuing intertropical desiccation which accompanied the late Pliocene expansion of continental ice sheets in North America added the final gloss to the late Cenozoic drying out of Australia's great inland rivers and lakes, which reached its full climax in the second half of the Quaternary (Figs. 28.5, 28.6, 28.7).

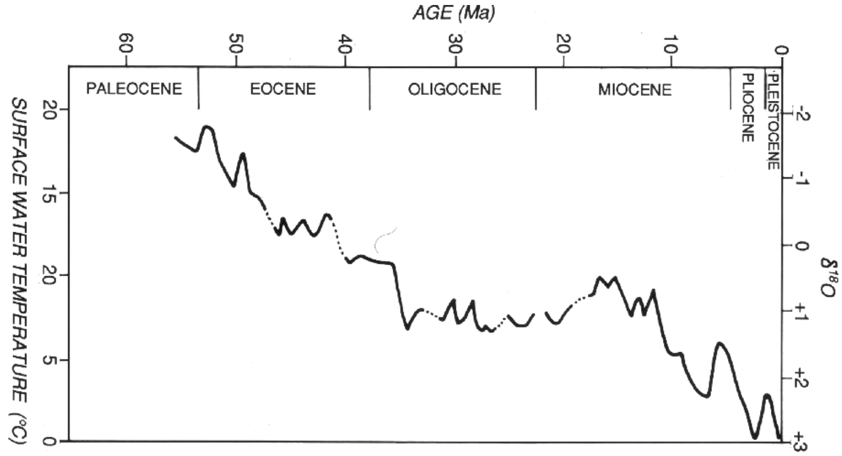

Fig. 28.4 Cenozoic sea surface temperatures in the Southern Ocean deduced from changes in oxygen isotopic composition of planktonic Foraminifera at DSDP sites 277, 279, and 281 (after Shackleton and Kennett 1975, Fig. 2)

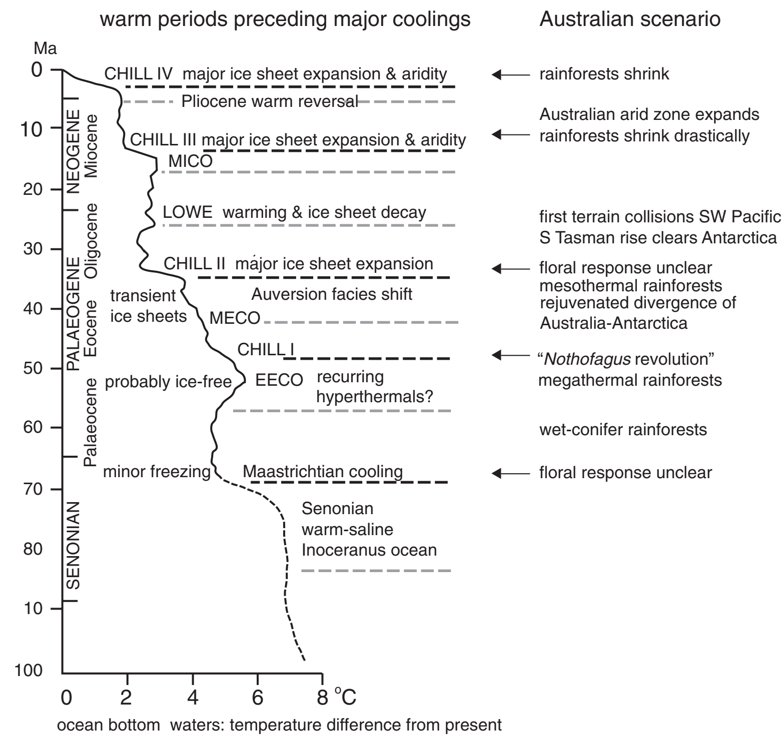

Fig. 28.5 Cenozoic events in Australia in global context (after McGowran et al. 2004, Fig. 2)

Quaternary Climatic Fluctuations

Recent palaeomagnetic, geomorphological, and geochemical investigations in the arid Amadeus Basin of central Australia have demonstrated that the onset of aridity did not become apparent until about the Jaramillo subchron (1.1 Ma) below the Brunhes–Matuyama palaeomagnetic boundary (Chen 1989). Before that time, from at least 5 Ma until about 0.9 Ma, the pattern of sedimentation in the Amadeus Basin was dominantly one of fluviolacustrine clay accumulation with intermittent drier intervals. Thereafter, quartz dune sands, groundwater evaporites, and wind-blown gypsum sands indicate an alternation between aridity and less arid intervals characterized by high regional groundwater levels and weak pedogenesis on stabilized dune surfaces (Chen 1989).

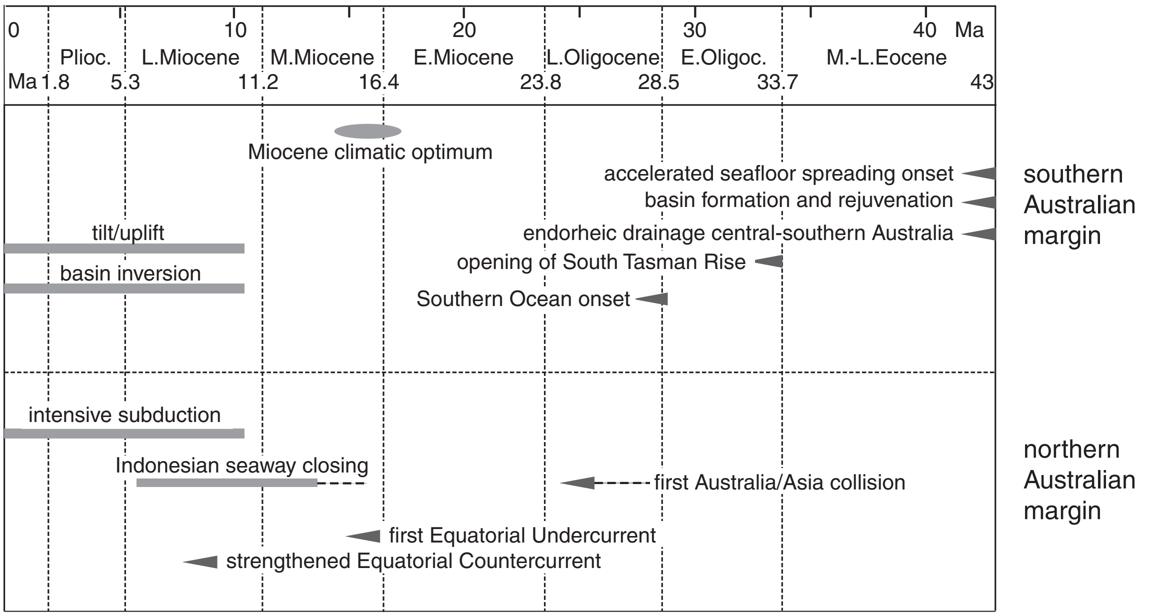

Fig. 28.6 Cenozoic environments of southern Australia (after McGowran et al. 2004, Fig. 9)

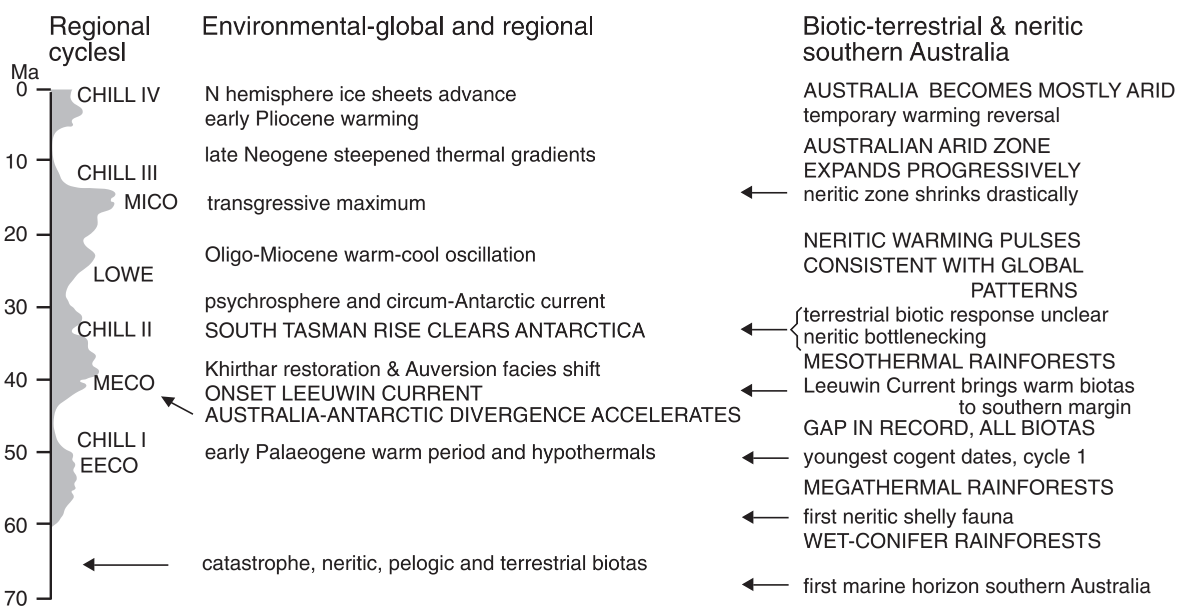

Fig. 28.7 Cenozoic events in northern and southern Australia (after McGowran et al. 2004, Fig. 28)

It is tempting to equate the change in regional climate evident in the change in depositional regime in central Australia to a change in global atmospheric circulation linked to the astronomically modulated cycles of changing global ice volumes and ocean temperatures. As a working hypothesis, we propose that aridity in both the Sahara and Australian deserts was more severe during the last 0.9 Ma than previously, and was associated with a change from the low-amplitude, high-frequency 41,000-year obliquity cycle, which dominated the late Pliocene and first half of the Pleistocene, to the high-amplitude but low-frequency 100,000-year orbital eccentricity cycle, which has dominated the last 0.9 Ma of the Pleistocene.

Desert dunes and sand plains occupy some 40% of mainland Australia and have long been a focus of attention. The relationship between dune orientation and present-day wind direction is well established. During and after the LGM dunes were active over 40% of the continent, including NE Tasmania. Preliminary OSL age estimates for dunes in the Strzelecki and Tirari deserts show well separated peaks at ~14 ka, 20 ka (LGM), and 34 ka, with minor peaks or shoulders at ~10.5 ka and 42 ka (Rhodes et al., 2004). The apparent trough between the 20 ka and 14 ka peaks suggests that few deposits have been preserved from that period. Either little dune building occurred then or subsequent reworking has removed the evidence. Dune sediments are best preserved if succeeded by a humid phase but have less chance of preservation if followed by an arid phase. Swezey (2001, 2003) makes a similar point in relation to the Saharan dunes. OSL ages for 20 aeolian dune samples from other locations in Australia show peaks at ~21 ka, 36 ka and 68 ka and a pronounced shoulder at ~43 ka. Combining all the data (54 age estimates) suggest that periods of dune-building and aeolian activity across Australia centred at ~14 ka, 20 ka, 35 ka, 42 ka and 70 ka. The event with the most samples is that at 20 ka when glaciers were at their maximum global extent and sea level at its lowest (LGM). Earlier TL ages (37 samples) by Nanson et al. (1992) lack the resolution of the new OSL data but do show broad peaks at ~30–40 ka and ~70 ka.

Hesse and McTainsh (2003) have investigated modern dust storms and Quaternary aeolian dust flux in marine cores. The marine record shows a threefold increase in dust flux during the LGM relative to the Holocene in temperate and tropical Australia. The causes appear to be weakened Australian monsoon rains (tropical north) and drier westerly circulation (temperate south). The Lake Eyre Basin/Simpson Desert and the Murray-Darling basin are the major present-day sources of dust. During 33–16 ka, there was enhanced aeolian dust flux to the east and south over the southern half of the continent. The northern limit of the dust plume was 350 km or 3° N of the present limit during 22–18 ka (Hesse et al., 2004). Williams and Nitschke (2005) have demonstrated that aeolian dust was a significant component of late Pleistocene valley-fills in and around the Flinders Ranges in semi-arid South Australia. These have been dated using paired OSL and AMS ^{14}C samples to ~33ka near the base and ~17 ka near the top of the sequence. Drawing upon over 200 ^{14}C, TL and OSL ages, Bowler (1998) and Bowler and Price, (1998) have established that aeolian dust (or *Wüstenquarz*) began to accumulate in the lunettes on the eastern side of Pleistocene Lake Mungo and adjacent lakes from ~35 ka until ~16 ka with a peak centred around the LGM. Clay dunes and gypseous lunettes were active on the downwind margins of seasonally fluctuating lakes in many parts of the SE and SW immediately before and during 21–19 ka. Major deflation of dry lake beds coincided broadly with the time of extreme aridity centred on the LGM (e.g., Lake Eyre: Magee and Miller, 1998).

Gingele and De Deckker (2005) have recorded aeolian dust in two cores that span the last 170 ka located on the continental margin of South Australia immediately south of Kangaroo Island in the Murray Canyons area. During periods of minimum insolation at this latitude, strong westerly winds blew dust from the continental interior, with peaks at ~20 ka, ~45 ka and ~70–74 ka. These periods are coeval with times of lake desiccation, dune building and sparse vegetation cover in the centre and south of Australia (Croke et al., 1996). They also note a conspicuous double ∂^{13}C minimum at 13 ka and at ~12–10.5 ka that reflects a major hydrographic change south of the Polar Front, during which Antarctic Intermediate Water was transferred to lower latitudes. These events may coincide with the Antarctic Cold Reversal and with the Younger Dryas.

In the lead up to the cold and arid Last Glacial Maximum there was a progressive weakening of the Australian summer monsoon. Johnson et al. (1999) analysed carbon isotopes in fossil emu eggshell from around Lake Eyre in central Australia. They found significant changes in the proportions of C4 grasses over the last 65 000 years. The data imply that the Australian monsoon was most effective between 65 and 45 ka, least effective during the LGM, and moderately effective during the Holocene. They noted that the effectiveness of the summer monsoon decreased at about the time that the megafauna became extinct and at about the time that humans arrived on the continent.

The evidence from lakes, lunettes and wetlands provides further insights into the late Quaternary evolution of the Australian desert. During the LGM the tropical northern lakes were mostly dry, except for Lake Carpentaria. Lake Eyre in central Australia was totally dry and its bed was actively lowered by wind erosion (Magee et al., 1995; Croke et al., 1996; Magee and Miller, 1998). Southeastern Australia shows cooler conditions but significant regional variations in hydrology. Lakes in the southwest and southeast were mostly dry or saline at this time. In the now arid to semiarid Flinders Ranges of South Australia a number of permanent wetlands occupied sheltered valleys within the mountains from at least 33 ka until somewhat after 16 ka (Williams et al., 2001). The persistence of such a wetland reflects the interactive influence of a number of factors: lower temperature and lower evaporation; low CO_2 levels and fewer trees, leading to rising groundwater levels; increased dust flux, trapped by the grass cover; reduced hillslope runoff and increased infiltration into the slope mantles, contributing to enhanced base flow; weak summer monsoon and fewer erosive events; lower cloud base and enhanced gentle winter rains (Chor et al., 2003).

Miller et al. (1997) used the temperature-dependent amino acid racemization reaction in radiocarbon-dated emu eggshells from the continental interior to reconstruct subtropical temperatures at low elevations over the last 45 ka. They concluded that millennial-scale average temperatures were at least 9°C lower between 45 and 16 ka than after 16 ka. There was a sharp change at 16 ka followed by rapid warming.

Impact of Key Tectonic and Climatic Events on Deserts

Throughout this chapter, we have been at pains to emphasize that many of the geomorphic attributes of our present-day deserts have a very ancient geological pedigree. We have stressed the role of Precambrian and early Phanerozoic tectonic inheritance in controlling the disposition of Mesozoic and Cenozoic uplands and lowlands in the Sahara. In Australia, we have seen that certain elements of the modern landscape are exhumed surfaces developed well before the final onset of Tertiary and Quaternary climatic fluctuations, and that some of these surfaces have been remarkably little altered since they were formed during the Precambrian. It is now appropriate to review some of the key tectonic and climatic events of the Mesozoic and Cenozoic which, directly or indirectly, had an important influence on the evolution of our modern deserts (Figs. 28.5, 28.6, 28.7).

Both hemispheres played a decisive role in the events which culminated in the late-Cenozoic desiccation of what are now the great tropical and temperate deserts of the world. Since Cenozoic intertropical desiccation was intimately associated with the cooling of Antarctica, we begin with a discussion of tectonic and climatic events in the Southern Hemisphere.

Initial separation of Australia from Antarctica started at 90 Ma and was fully effective by 45–50 Ma (Fig. 28.5). Later opening of the Drake Passage between South America and Antarctica towards 30–25 Ma resulted in the establishment of a circum-Antarctic ocean current (Fig. 28.5). Antarctica was now thermally isolated from warmer ocean waters to the north, and rapid cooling ensued. In the Southern Ocean, the changing isotopic composition of both planktonic and benthic Foraminifera is evidence of a dramatic cooling of deep ocean water as well as surface water. Cumulative ice build-up in Antarctica saw the creation of mountain glaciers followed by the growth of a major ice cap, first in East Antarctica and later in West Antarctica (Fig. 28.6). Australia, meanwhile, was moving north into dry subtropical latitudes. Within Australia, forest gave way to woodland, and woodland gave way to savanna. The net effects for Australia were climatic desiccation, progressive disruption of the drainage network, expansion of the desert, and successive plant and animal extinctions (Fig. 28.7). We turn now to the Northern Hemisphere.

Collision between Greater India and Asia extends back to about 45 Ma, and by 20–15 Ma the resulting underthrusting of Greater India beneath Asia had resulted in early uplift of the Himalayas. This uplift continued at least intermittently during the Miocene, Pliocene and Quaternary and continues to this day. Development of the Indian monsoon during the late

Miocene (or earlier) was one climatic outcome of this uplift. Another was the genesis of the easterly jet stream emanating from the Tibetan Plateau.

In East Africa, uplift and rifting created the Neogene depocentres with their unrivalled record of Pliocene and Pleistocene hominid evolution. The emergence of *Australopithecus afarensis*, a bipedal but small-brained early Pliocene hominid, may well be linked to the 6–5-Ma Messinian salinity crisis which led to the genetic isolation of Africa from Eurasia. During this terminal Miocene event, the Mediterranean Sea dried out completely, refilled and dried out again on about a dozen occasions, resulting in the accumulation of evaporites and anhydrite deposits up to a kilometre thick, representing roughly 6% of the total oceanic salt supply. The precursors to this 'salinity crisis' were the Miocene and earlier shrinking of the Tethys Sea (with an associated change to a more seasonal rainfall regime with lower annual precipitation) and the more immediate late Miocene glacio-eustatic drop in sea level caused by final accumulation of the West Antarctic ice cap. It was the 40-m fall in global sea level in the very late Miocene that exposed the Gibraltar sill separating the Atlantic from the Mediterranean, so that the Mediterranean became a closed basin in which evaporation greatly exceeded inputs from coastal rivers and local rainfall. The result was extreme aridity along the North African littoral, together with fluvial downcutting by the late Miocene Nile to carve a gorge over 1000 km long and up to 2 km deep at its northern end.

Late Pliocene cooling and desiccation towards 2.5 Ma (Fig. 28.5) caused the tropical lakes of the Sahara and East Africa to shrink, and in upland areas such as Ethiopia was associated with expansion of grassland at the expense of montane forest. It may be no coincidence that the first stone tools made to a replicated pattern make their first appearance at about this time, together with an increase in meat-eating by our hominid ancestors.

Northern polar cooling towards 15 Ma followed break-up of Laurasia. Antarctic cooling and ice build-up at about this time triggered a major change in ocean bottom water circulation (Fig. 28.6). Pliocene closure of the Panama isthmus was indirectly responsible for enhanced snowfall in high northern latitudes, reflected in Arctic ice accumulation by 3.5 Ma and a rapid expansion of North American ice caps at 2.5 Ma. Initiation of the Laurentide ice sheet ushered in the late Pliocene and Quaternary glacial–interglacial cycles, with consequent glacio-eustatic sea-level oscillations and a global pattern of spatially and temporally alternating morphogenetic systems. Glacial maxima were times of desert expansion and forest retreat. During at least the last 0.7 Ma, each cycle was roughly 100 ka long and modulated by the 100-ka orbital eccentricity cycle.

The net effects of the late Cenozoic changes in tectonics, climate, and ocean circulation were therefore a global increase in latitudinal temperature gradients and in the seasonal incidence of rainfall, final evolution of the incipient deserts into regions of scanty and unreliable rainfall, and the emergence in Africa of upright-walking, stone toolmaking ancestral humans who later discovered how to make and use fire, and ultimately occupied every continent except Antarctica.

The story of the Cenozoic tectonic and climatic events which culminated in the slow emergence of the world's great deserts is thus also, in a very real sense, the story of the global environmental changes which were associated with the origins of the first humans in Africa some 2.5 Ma ago. From that time on, humans began to modify their habitats, sometimes contributing unwittingly to desertification or local expansion of the desert environment. Notwithstanding our destructive proclivities, the great deserts would be where they are even if humans had never appeared on the face of this Earth. We must needs learn to live with our Cenozoic inheritance.

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