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…1, and 5.6), J. F. Thackeray (Fig. 6.7), R. G. Klein (Fig. 6.8), and H. J. Deacon (Fig. 8.1).

The initial synthesis of palaeoclimatic evidence by NL was carried out at the Univesity of Cape Town, where Professor Richard Fuggle, of the Department of Environmental and Geographical Science, gave valuable support and encouragement for the project. Logistic and institutional support for the final phases of NL's work on the book was provided by the Department of Geology, Arizona State University. JD acknowledges with grateful thanks the continued support of the Department of Archaeology at the University of Stellenbosch. The Institute for Cartography at the University of Stellenbosch was responsible for drawing all the figures except 5.2 (J. Lancaster), 6.2 (V. B. Geleijnse), 3.3, 6.3, 6.4, 6.6, 6.7 and 6.9–6.13 (J. Deacon), and 7.1–7.3 (N. Lancaster).

We have enjoyed stimulating discussions on the Late Quaternary environments of Southern Africa with many colleagues, several of whom have read and commented on sections of this book. They include D. M. Avery, P. B. Beaumont, K. W. Butzer, H. J. Deacon, J. R. Grindley, R. G. Klein, R. R. Maud, J. E. Parkington, T. C. Partridge, J. Rogers, U. Rust, A. Scholtz, L. Scott, P. Shaw, A. S. Talma, J. T. Teller, J. F. Thackeray, M. L. Tusenius, P. D. Tyson, E. M. van Zinderen Bakker, J. C. Vogel, and J. D. Ward. Helen Barlow and Judith Lancaster helped considerably with proof-reading and other tasks in the preparation of the manuscript, and we thank them most sincerely.

Finally, we should like to thank our spouses, H. J. Deacon and Judith Lancaster for their assistance and encouragement at all stages of the work. Any flaws or faults can only be our own.

1987

J. D.
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Deacon, J. and Lancaster, N. 1988.
Late Quaternary palaeoenvironments of southern Africa.
Clarendon Press, Oxford.

Contents

1 Introduction

1.1 Approaches to Late Quaternary research

The main aim of palaeoclimatic research is to describe variations in climate in the period beyond the range of the meteorological record. The ultimate goal is theoretical: to understand the physical causes of these variations. By applying principles established through modern climatological research, it is possible to understand the mechanisms of change and the pattern of palaeoclimates in time and space. This has considerable practical significance in, for example, studies of the effects of a possible future increase in atmospheric carbon dioxide content. Palaeoclimatic data provide the best means of testing models designed to assess future climatic trends, so there is an important feedback between the present and the past, and between the past and the future. Palaeoclimatic data are also the only means of studying ecosystem history.

Geographically, this book focuses on Southern Africa, i.e. Africa south of the Zambezi River. It therefore includes a wide range of environments and habitats, excluding only tropical forest and glaciated highlands. We assess the current status of Late Quaternary palaeoclimatic studies in this region and set out some of the priorities for future research. In Southern Africa, the Quaternary has been studied for the most part by two distinct groups of researchers: those aligned to the geological sciences who are interested in landforms and sedimentary deposits, and the commercial and non-commercial aspects of soil types and formation; and those who are concerned with archaeology and the prehistory of human settlement in Southern Africa. Both need to discover how old the deposits are, and under what climatic and other environmental conditions they formed.

The Quaternary includes the Pleistocene and Holocene epochs. The 'official' Pliocene–Pleistocene boundary dates to 1.6 Ma at Vrica in Italy (Backman *et al.* 1983) and the Pleistocene–Holocene boundary is placed at 10 000±250 years ago (Fairbridge 1983, p. 238). Within the Pleistocene, the Lower–Middle Pleistocene boundary coincides with the Brunhes–Matuyama palaeomagnetic boundary at 700 000 years ago and the Middle–Upper Pleistocene boundary coincides with the oxygen isotope deep sea core stage 6–5 boundary at about 130–125 000 years ago (Fairbridge 1983; Turon 1984). These dates do not mark periods of

2 Present and past climates of Southern Africa

2.1 Present day climates

Today, Southern Africa is dominated by dry climates and strongly seasonal precipitation regimes. This is the datum against which all Late Quaternary climatic changes must be viewed. Detailed descriptions and analyses of the climates of the subcontinent are contained in Jackson and Tyson (1971), Schulze (1972, 1984), and Tyson (1986), to which the reader is referred for further information. This section outlines the main features of Southern African climates as a background to subsequent discussions of palaeoclimates and palaeoclimatic models.

The climates of Southern Africa are strongly influenced by the latitudinal position of the subcontinent, which tapers southwards from 18 to 35° south. The west coast is paralleled by the cold northward flowing Benguela current, whilst the warm south flowing Mozambique and Agulhas currents lie off the east and south-east coasts. Many areas of the subcontinent lie at elevations of over 1000 m. This, together with the narrow width of the land mass, exerts a moderating influence on the climate (Schulze 1972).

Southern Africa lies almost entirely within the subtropical high pressure belt which is separated by the land mass of Africa into two cells: the South Atlantic and Indian Ocean anticyclones. The mean circulation over Southern Africa is therefore anticyclonic throughout the year above the atmospheric boundary layer (Tyson 1986). North of 20° south, climates are strongly influenced by the seasonal movement of the Intertropical Convergence Zone (ITCZ). The circumpolar westerlies and their associated temperate low pressure disturbances lie to the south of the subcontinent. Southern Africa therefore lies in a zone of interaction between tropical and temperate circulations which influences climates at all seasons. yes!

2.1.1 Circulation patterns

The South Atlantic and Indian Ocean subtropical anticyclonic cells are centred on 30° south, with 5–6° of seasonal latitudinal movement. The

Indian Ocean high tends to move south-eastwards in summer, whilst the South Atlantic cell weakens and moves north-west in winter. In summer a weak thermal low pressure cell may develop over the central interior of the continent and is linked by a trough to a tropical low north of Botswana (Taljaard 1981). The position of this low varies longitudinally along the southern branch of the ITCZ in the region of the Zaire Air Boundary and occasionally moves south to northern Botswana and Namibia. There is also a weak ridge of high pressure over the northern Cape Province. In winter, the mean circulation at the surface of the Southern African plateau is strongly anticyclonic, with a single large high pressure cell developed over the north-eastern Transvaal. The intensification and equatorwards extension of the westerlies reaches a maximum at this time of the year. The change to a single high pressure cell occurs in March, giving rise to an influx of moist tropical air over the western parts of Southern Africa and an autumnal rainfall maximum in these areas.

2.1.2 Climatic elements

Rainfall

Figure 2.1. shows the distribution of annual precipitation over Southern Africa. Rainfall is at a maximum along the eastern escarpment zone and decreases westwards. A second rainfall maximum occurs in the south-western Cape. The orographic effects of the Drakensberg escarpment and the Cape Fold Belt are very marked. In the south-western Cape, rainfall increases from 400 mm on the Cape Flats to over 2000 mm in the adjacent mountain ranges. The lowest rainfall totals are recorded along the Namib coast where less than 20 mm falls per year. The 500 mm isohyet divides Southern Africa into a wetter eastern and a dry western half. dry west wetter east

Precipitation over the northern and eastern parts of Southern Africa occurs mostly in summer and more than 80 per cent of rainfall in these areas occurs between October and March. In the south-western Cape, over 80 per cent of rainfall occurs in the winter months of April to September. The contribution of winter rainfall to the annual total decreases rapidly north-eastwards. Only the narrow southern Cape coastal region and the adjacent mountain ranges receive all season rainfall, with equinoctial maxima. seasonal rain

Rainfall over the summer rainfall zone is mainly convectional in origin. Conditions favouring the development of convectional cells are controlled not only by diurnal heating and mesoscale forcing, but also by synoptic conditions which give rise to strong uplift of unstable moist air masses. A common synoptic scenario for the development of large scale convective activity is the occurrence of a trough in the tropical easterly convectional rain

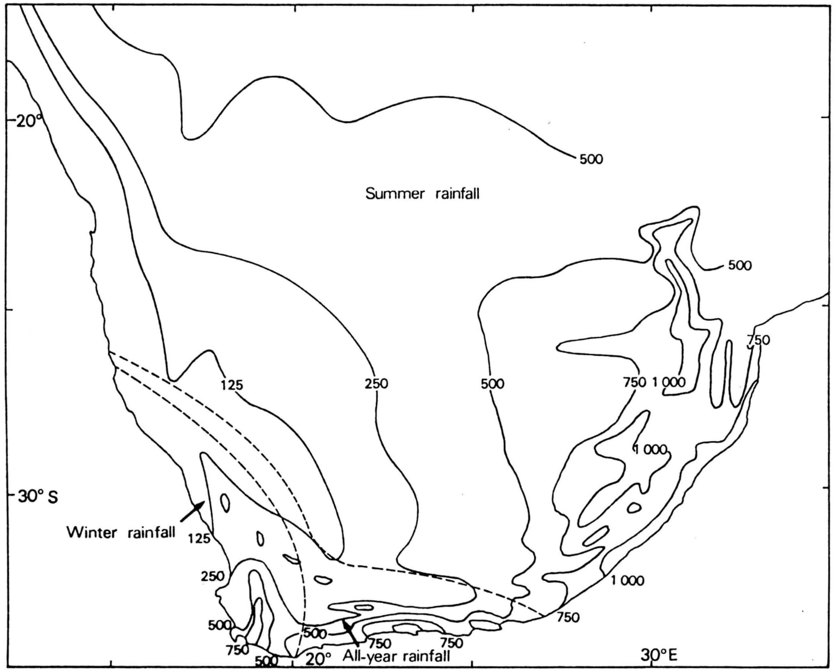

Fig. 2.1 Simplified mean annual rainfall (mm) for Southern Africa (after Schulze 1972).

circulation (Taljaard 1981). The passage of frontal systems also may trigger storms, but frequently no single synoptic scale forcing mechanism may be identified (Tyson 1986). In the winter rainfall zone, most rainfall is associated with the passage of cold fronts.

fine, dry weather

Periods of fine and dry weather over Southern Africa are associated with the development of large anticyclonic cells over the subcontinent with their attendant strong subsidence and temperature inversions. Such conditions are most common in winter, but they may also occur in summer. If they persist for extended periods at this season, heat waves and severe desiccation are the result (Jackson 1961).

variability

wet + dry cycles

Rainfall totals vary from year to year, with the interannual variability rising from 20 per cent over the wetter eastern areas of Southern Africa, to 60 per cent in the south-western Kalahari and more than 80 per cent in the central Namib. Tyson (1986) has identified a clear spatial pattern of extreme wet years (rainfall exceeding 125 per cent of normal), which are more common in the drier parts of the region and extreme dry years (rainfall less than 75 per cent of normal) which are more frequent in wetter areas. In the summer rainfall area, Tyson and his co-workers

(Tyson *et al.* 1975; Tyson and Dyer 1975, 1978; Tyson 1986) have recognized a weak 18-year periodicity in rainfall totals, comprising 9-year spells of predominantly wet years, followed by nine predominantly dry years. There is a strong spatial patterning to rainfall periodicities (Tyson 1986) with a 10–12-year oscillation being noted in the southern Cape. Over north-western parts of South Africa and Namibia, the oscillation is a 6-yearly one. The west central interior shows a quasi biennial cycle.

Surface temperatures

hi altitude ocean infl.

Compared to areas at similar latitudes (for example, North Africa and Australia), Southern Africa experiences relatively low temperatures as a result of the general elevation of the subcontinent and its proximity to oceanic influences. The range of temperatures (Fig. 2.2) is considerable and local topographic effects are important. The highest temperatures are recorded from the southern Kalahari, Mozambique, and western Zimbabwe where mean annual temperatures reach 20–24°C. The lowest

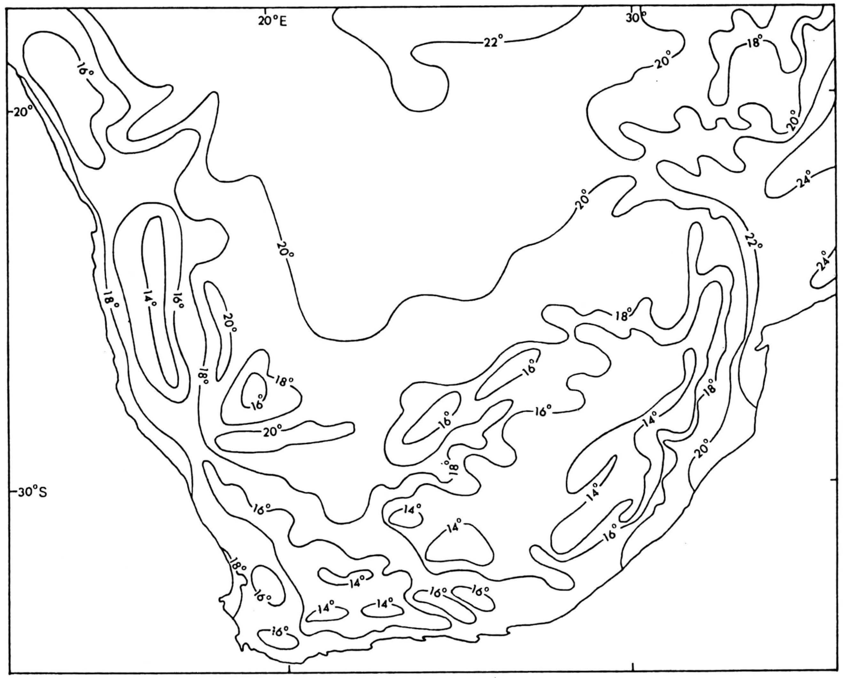

Fig. 2.2 Simplified mean annual temperature (°C) for Southern Africa (after Schulze 1972).

temperatures (11–12°C mean annual temperature) occur in the eastern highlands and Cape mountains. The importance of oceanic effect is shown by a comparison between the mean annual temperatures at 29° south, which are 15°C on the west coast and 21°C on the east coast. The greatest diurnal temperature ranges are found in the interior, especially in the Kalahari and Karoo. The highest frequency of temperatures below freezing occurs along the Natal–Lesotho mountains and escarpment zones where there may be more than 90 days a year when the temperature falls below 0°C.

Humidity and evaporation

The major moisture source for most of Southern Africa is the Indian Ocean and there is a strong east to west gradient in atmospheric moisture content, especially in summer (Jackson 1961). Relative humidity at 14.00 hours declines from 70–80 per cent in coastal areas to 50 per cent in eastern plateau areas, and less than 30 per cent in the west (Schulze 1972).

Pan evaporation rates increase inland and from east to west, reaching a maximum of 4000 mm per year in the south-western Kalahari. They are least along the southern Cape coast and in the eastern highlands.

Surface winds

The pattern of surface winds is strongly influenced by the generally anticyclonic nature of circulation patterns in Southern Africa. In the northern interior, winds are easterly to north-easterly and strongest in summer. Northerly to north-westerly winds prevail all year in the central interior. In the western interior, there is a strongly seasonal pattern of winds which are north-north-east in winter, but south-south-west in summer. Similarly, in the eastern and south-eastern interior, there is a monsoonal effect with north-westerly winds in winter, but south-easterly winds in summer (Schulze 1972).

Winds in coastal areas are generally stronger than in inland areas, especially in the Cape Province. They generally parallel the coast. On the south and east coasts winds are southerly to south-easterly onshore in summer and northerly or north-westerly offshore in winter. In the south-western and western Cape, winds are south-easterly or southerly in summer, but north-westerly or westerly in winter.

Causes of rainfall variations

The synoptic conditions which give rise to extended wet or dry spells have been discussed by Harrison (1986) and Tyson (1984, 1986). In the summer rainfall zone, wetter spells on the scale of days, seasons and years are associated with the development of lowered pressure over Southern Africa and increases in pressure over the South Atlantic. In dry spells, the converse applies. The development of such circulation patterns appears to be linked to variations in the zonal Walker circulation, the so-called Southern Oscillation. In the high phase of the Oscillation, tropical easterly flow strengthens and the westerlies contract polewards, but become stronger around latitude 47° south. During the dry (low) phase, the westerlies weaken polewards, but expand and strengthen northwards and the tropical easterlies weaken (Tyson 1986). Meridional circulations over Southern Africa and the adjacent oceans, together with interactions between tropical and temperate perturbations, provide the link between the African Walker cell and the subtropical circulations. The major form of interaction between tropical easterly and temperate westerly circulations occurs in the summer months and takes the form of extended cloud bands, visible on satellite images, which link tropical easterly waves and temperate westerly waves and low pressure systems. Enhanced convective activity and rainfall over the interior of Southern Africa is usually associated with the cloud bands.

On the basis of these observations, Tyson and his co-workers (Tyson 1986) have put forward a model for the atmospheric circulation in extended wet and dry spells. Most rainfall over the interior of the subcontinent results from the favourable conjunction of deep tropical and westerly disturbances, which allow the formation of major north-west–south-east orientated cloud bands. During wet spells (Fig. 2.3), the ascending part of the African Walker cell is located over tropical Africa. The subtropical jet stream is strengthened to the east of the subcontinent. The ITCZ is also strengthened, resulting in more easterly waves and enhanced easterly flow and advection of moisture from the Indian Ocean. Changes in the amplitude of the upper level Atlantic wave result in the preferred location of the cloud bands over Southern Africa. At the same time, there is a southward shift of storm tracks because of higher poleward energy fluxes and a lower meridional temperature gradient, resulting in drier winters in the winter rainfall zone.

In dry spells (Fig. 2.3) the rising limb of the African Walker circulation is situated off the coast of Africa, the ITCZ is weakened and tropical easterly waves diminish. Pressure rises over Southern Africa and moisture advection to the subcontinent decreases. The South Atlantic anticyclone is weaker and the Atlantic wave diminishes in amplitude, such that cloud bands locate preferentially over the Indian Ocean. Poleward energy fluxes decrease, but temperature gradients increase, leading to an equatorwards movement of storm tracks, bringing more winter rain to the south-western Cape.

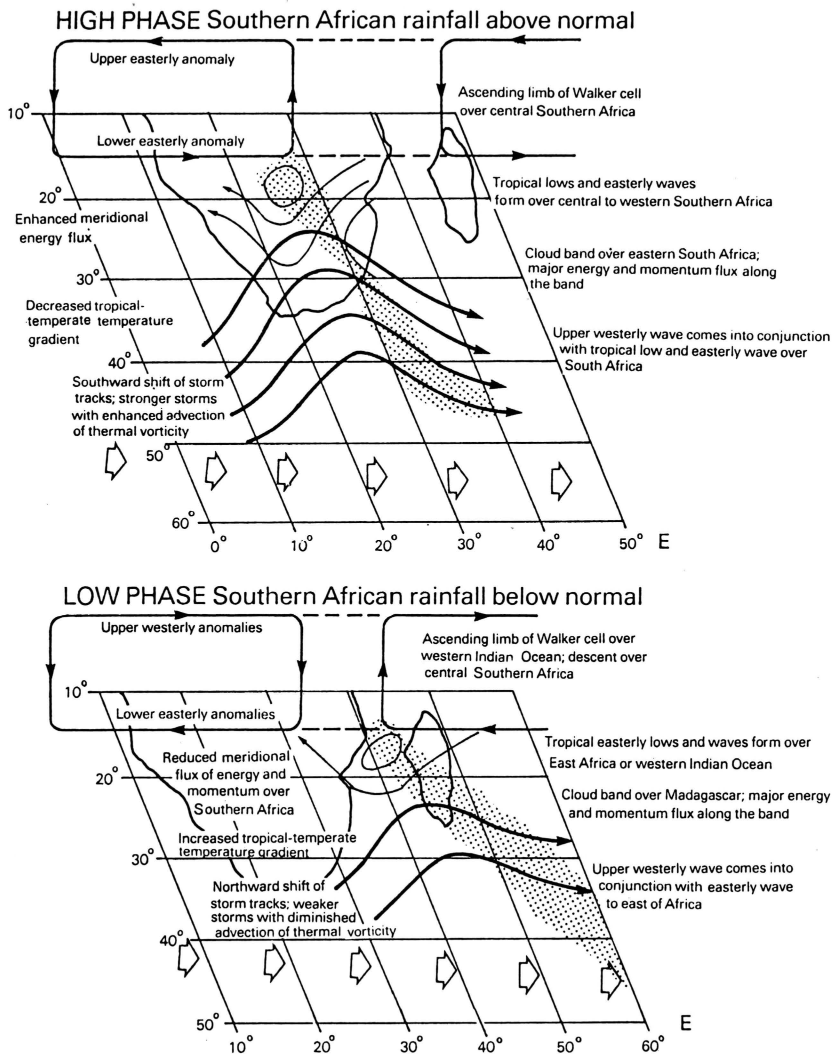

Fig. 2.3 Schematic representation of the anomalous Walker circulaton over Southern Africa during high phase of the Southern Oscillation (above) resulting in above-normal summer rainfall; and during low phase (below) resulting in below normal summer rainfall and above normal winter rainfall (after Tyson 1986, fig. 9.4).

2.2 Palaeoclimatic models

Global circulation models have been calculated for the Last Glacial Maximum and in this section the results will be compared with

atmospheric circulation models that have been proposed for Southern Africa. The latter can be usefully divided into models that assume:

(a) an equatorward movement of circulation belts in both hemispheres;
(b) stronger circulation patterns without latitudinal displacement; and
(c) a longitudinal displacement of circulation patterns.

All three tend to focus on a few features of atmospheric circulation rather than taking into account all the possible variables, but this has been a necessary device for testing expectations against the field data. In this section we focus on the explanations proposed for Last Glacial Maximum conditions.

2.2.1 Global circulation models for the Last Glacial Maximum

By the early 1970s sufficient data had been accumulated from deep sea cores to allow some correlation between these and modern climatic data gathered around the world. In 1971, the influential CLIMAP (Climate, Long-range Investigation, Mapping, and Prediction) programme was launched to reconstruct from quantitative geological evidence the average state of climatic boundary conditions (sea-surface temperature, ice extent, ice elevation, continental albedo, wind velocity, rainfall/precipitation, land surface temperature, etc.) for the Last Glacial Maximum at 18 000 BP, and to measure the oscillations of Pleistocene climate worldwide. As part of this programme, general circulation models (GCMs) were developed which mathematically simulate ice age climatic conditions by explaining in dynamic terms the complex of processes which maintain a particular climate in equilibrium (CLIMAP 1976, p 1131).

GCMs are computed from equations governing the transfer of heat and momentum in the atmosphere based on specifications for individual locations of incoming solar radiation, land topography, sea surface temperature, surface albedo, and ice sheet extents and elevations. These models can then be compared against the evidence gathered from biological and geological data. The general acceptability of the results of such a simulation are shown by Heath (1979, fig. 1) who contrasts a simulated model for present-day conditions and a map summarizing observed present-day precipitation around the world. The correspondence is reasonable, but not absolute. Simulated models of boundary conditions for the Last Glacial Maximum have been computed for global sea-surface temperature (CLIMAP 1976), land surface temperature, surface heat balance, sea level pressure (Gates 1976; Manabe and Hahn 1977), precipitation (Heath 1979), relative humidity, cloudiness, evaporation rate, evaporation patterns, and wind velocities (Gates 1976;

Sarnthein 1978). In a comparison of five simulation models, Williams (1978) found considerable variation in temperature estimates for particular localities, but a general agreement that the Last Glacial Maximum was cooler and distinctly drier worldwide. These generalizations provide a broad standard against which field observations can be compared. Some of the GCMs for Africa are summarized in Fig. 2.4. Together, the CLIMAP data and GCMs confirm that 18 000 years ago temperatures were markedly cooler and that precipitation was generally lower. What is still at issue is just how much cooler and how much drier (if at all) it was at individual stations in Southern Africa. In order to assess this problem we review models proposed in the past to explain how and why Last Glacial Maximum climates differed from those of the present.

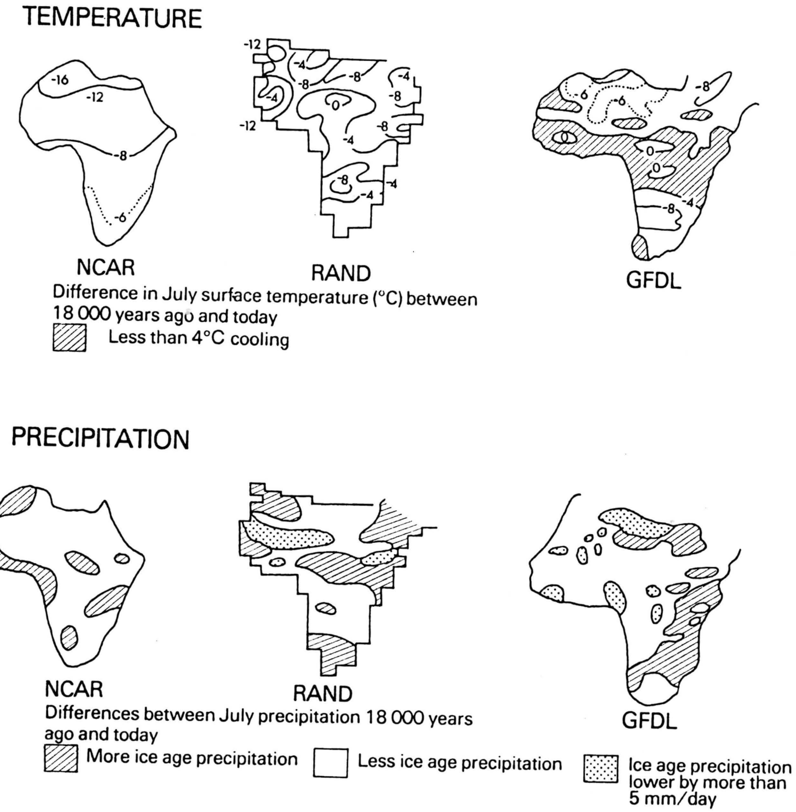

Fig. 2.4 Comparisons between temperature and precipitation estimates from three GCMs (General Circulation Models) for Africa at 18 000 BP. NCAR after Williams *et al.* (1974), RAND after Gates (1976), and GFDL after Manabe and Hahn (1977).

2.2.2 Models for Last Glacial Maximum climates in Southern Africa

It is inequalities in temperature which provide the energy for air circulation and, therefore, climate. The presence or absence of a polar ice cap on and around Antarctica has a significant effect on the southern hemisphere air and ocean temperatures, as well as on the wind and ocean currents, as can be shown in the scale of global climatic changes that took place before and after Antarctica moved into position over the South Pole (Van Zinderen Bakker 1976; H. J. Deacon 1983; Flohn 1984).

The Antarctic ice sheet at present contains 90 per cent of the world's ice and covers an area of 14 million km². In addition, the Antarctic sea-ice varies from 4 million km² in March, to 20 million km² in September. By contrast, the Arctic sea-ice cover has an extent of only 7–12 million km² in September and March/April, respectively. Thus, while the heat budget of the Arctic leads to minimum temperatures of −35°C, those in the Antarctic drop as low as −70° or lower (Pittock 1978; p 4), leading to a pole-to-equator temperature gradient which is 40 per cent larger over the southern hemisphere than over the northern hemisphere (Lamb 1972, p. 94). This anomaly results in mid-latitude westerly winds in the southern hemisphere being much stronger than those in the northern hemisphere. However, during a glacial maximum, this anomaly would be partly adjusted because of the much larger extension of ice over the northern continental landmasses, with a consequent increase in the northern hemisphere pole-to-equator temperature gradient. The strength of the present-day southern hemisphere westerlies leads to a displacement of the ITCZ 7° of latitude northward of the geographical equator in Africa (Flohn 1967, 1984; Pittock 1978).

The positioning and strength of the southern hemisphere westerlies are important aspects in the assessment of glacial maximum conditions for Southern Africa because they bring much of the winter rainfall to the western and southern Cape at present, and because strong westerlies help to maintain the zonality of the present-day circulation pattern by preventing the poleward migration of subtropical anticyclones (Boucher 1975, p. 280). The zone between the pack ice and the equatorward boundary of the westerlies, i.e. between the circumpolar trough and the subtropical ridge is dominated by cyclonic circulation, while anticyclones prevail to the north of the subtropical ridge. These two circulation patterns at present maintain the summer rainfall maximum in the interior, eastern and southern margins of Southern Africa on the one hand and the winter rainfall maximum on the western and southern margins on the other.

Initial interest in reconstructing past climatic changes in Southern

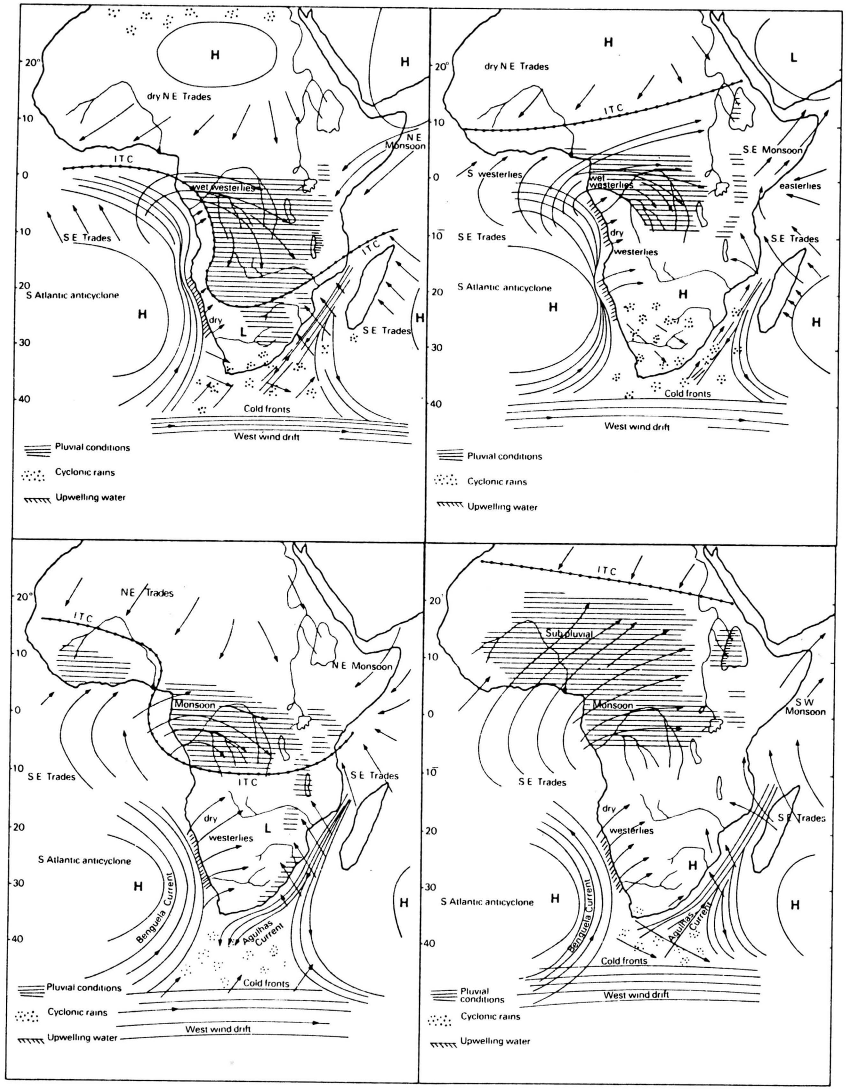

Fig. 2.5 Model of some palaeoclimatic trends in Africa proposed by Van Zinderen Bakker (1967). Top left: glacial southern summer; Top right: glacial southern winter; Bottom left: interglacial southern summer; Bottom right: interglacial southern winter. (Reproduced with permission from the author and The University of Chicago Press. ©1967 Wenner-Gren Foundation.)

model 1.
V. Zind Bakk

Africa was stimulated mostly by the palynological work of Van Zinderen Bakker (1957, 1962*a*). His first comprehensive reconstruction of glacial/interglacial climates considered the African continent as a whole (Van Zinderen Bakker 1962*b*, 1967) and maps of the major circulation patterns were drawn up (Fig. 2.5). He based these on the assumption that during the Last Glacial Maximum the zonal climatic control on the latitudinal positions of the major circulation patterns was in general the same as at present, but because the polar high pressure cells extended further towards the equator, the climatic belts were assumed to have contracted, accelerating the wind and ocean circulation systems. This resulted in a southward displacement of the ITCZ. One of the major implications was that the winter rainfall region of the western and south-western Cape would have been shifted further north than at present (Van Zinderen Bakker 1976, p. 144), with the corollary that the vegetation typical of the Cape Ecozone would have extended northwards into South West Africa, Namibia, and into the Cape and Orange Free State north of the Cape Fold Mountains.

Reiterating this model some years later, Van Zinderen Bakker (1976) (Fig. 2.6) cited confirmation for the hypothesized northward shift of the climatic belts in evidence for grassland habitats in the southern Cape, for archaeological proxy evidence which indicated that many sites in the southern Cape had no occupation deposits dating to the Last Glacial Maximum, and for geomorphological evidence which indicated much cooler temperatures in the Drakensberg. Data were also available at that

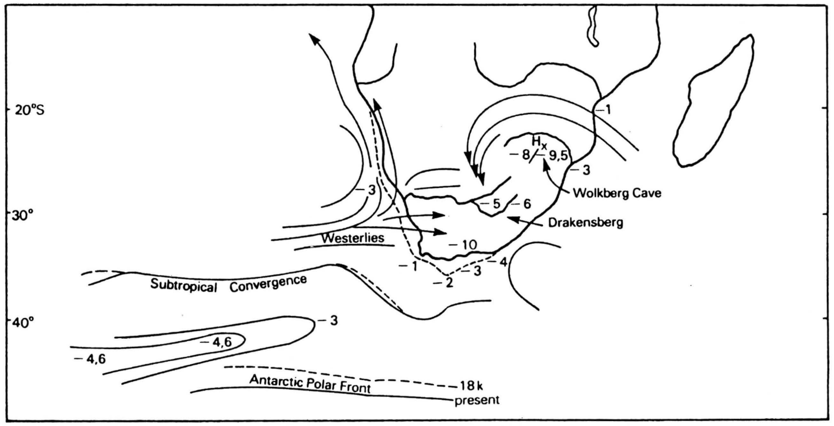

Fig. 2.6 Model of palaeoclimatic trends in Southern Africa at *c.* 18 000 BP from Van Zinderen Bakker (1982*a*). Ocean temperature anomalies after Newell *et al.* (1981). Position of subtropical convergence and Antarctic Polar Front after Morley and Hays (1979). Kalahari wind directions for summer and winter. (Reprinted from: *Palaeoecology of Africa*, (ed. J. C. Vogel, E. A. Voigt, and T. C. Partridge) Volume 15. Southern African Society for Quaternary Research, 1982. A. A. Balkema, P.O. Box 1675, Rotterdam, Netherlands.)

time to show that ocean temperatures were between 2 and 5°C cooler than at present off the Southern African coasts, and the palaeotemperature record of a stalagmite from the central Transvaal indicated a glacial maximum temperature of 8–9.5°C lower than at present (Talma *et al.* 1974). Van Zinderen Bakker (1976, p. 169) proposed, then, that glacial maxima were marked by dry and cool conditions in the north and east of Southern Africa, and cold temperatures, strong winds, and wet winters in the southern section. This hypothesis was of considerable value in giving a scenario for glacial and interglacial climates that could be tested in the field (H. J. Deacon 1983).

model 2

In a more detailed model that focused on Africa north of the equator, Nicholson and Flohn (1980) suggested an equatorward displacement of circulation features in the southern hemisphere that would have led to higher rainfall during the Last Glacial Maximum in the region between about 25 and 30° south, drier conditions to the north and about the same rainfall as today in the extreme south (Fig. 2.7).

model 3

Another model, proposed by Butzer *et al.* (1978*b*) stresses stronger circulation. It suggests that at glacial maxima, with the westerlies positioned at much the same latitudes as at present and an ITCZ southward of its present position, warm moist tropical air from the Congo and the Indian Ocean would have been advected more often over the summer rainfall region bringing higher rainfall, whereas the winter rainfall region would have received less rain except perhaps for the extreme south-west. A variation proposed by Van Zinderen Bakker (1980, 1982*a*) and Heine (1982; Fig. 2.8) also assumes strengthening of the circulation pattern, but with more vigorous trade winds bringing winter rains as far north as the southern Kalahari, and summer rainfall over the Kalahari, but not as far west as Namibia.

model 3

The third model invokes longitudinal instead of latitudinal displacement of high pressure cells (Tyson 1986). In examining modern weather patterns, Dyer (1976) showed a relationship between annual rainfall and the position of the subtropical highs over the South Atlantic and South Indian oceans, and Cockcroft *et al.* (1987) have used the observations to hypothesize how this would have affected Last Glacial climates as well. In a similar study in Australia (Pittock 1973), 84 per cent of the total variance in rainfall could be accounted for when the mean latitude and longitude for the high pressure belts was given. With the drought in the summer rainfall area in 1982–83 the focus has fallen on the effects of the Southern Oscillation or the related 'El Nino' effect in which the Southern Indian ocean high pressure cell is displaced northeastwards of its 'normal' position bringing with it much drier conditions in the summer rainfall region of Southern Africa. The implication is that such conditions may have prevailed over extended periods. Pittock and Salinger (1982) have proposed that just such an extended Southern

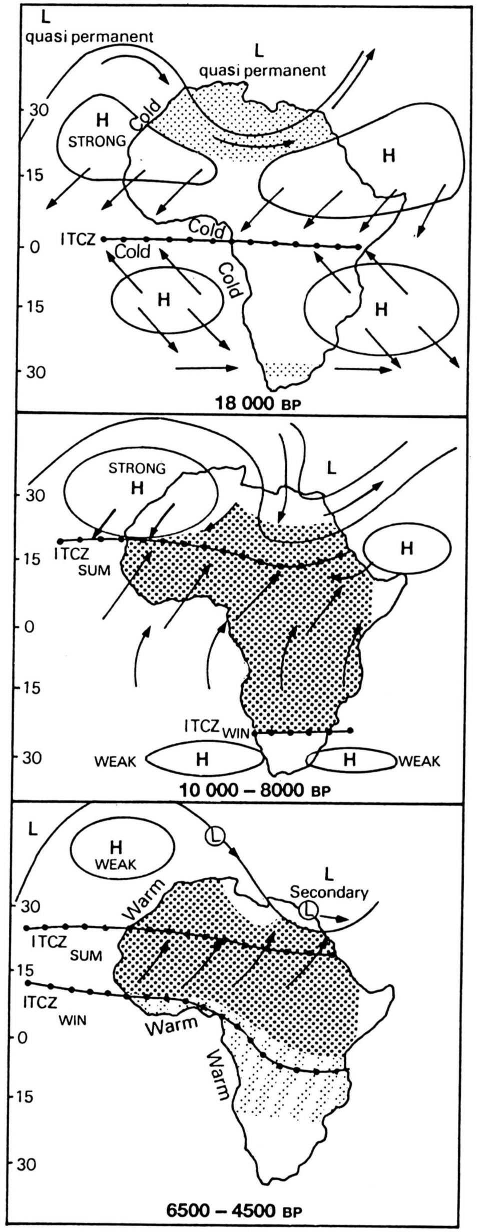

Fig. 2.7 Atmospheric circulation model for *c.* 18 000 BP (above), *c.* 10 000–8000 BP (middle) and *c.* 6500–4500 BP (below) proposed by Nicholson and Flohn (1980). Dark shading: more humid than present; light shading: drier than present; hatching: drier than previously, but more humid than present. (Copyright ©1980 D. Reidel Publishing Company. Reprinted by permission.)

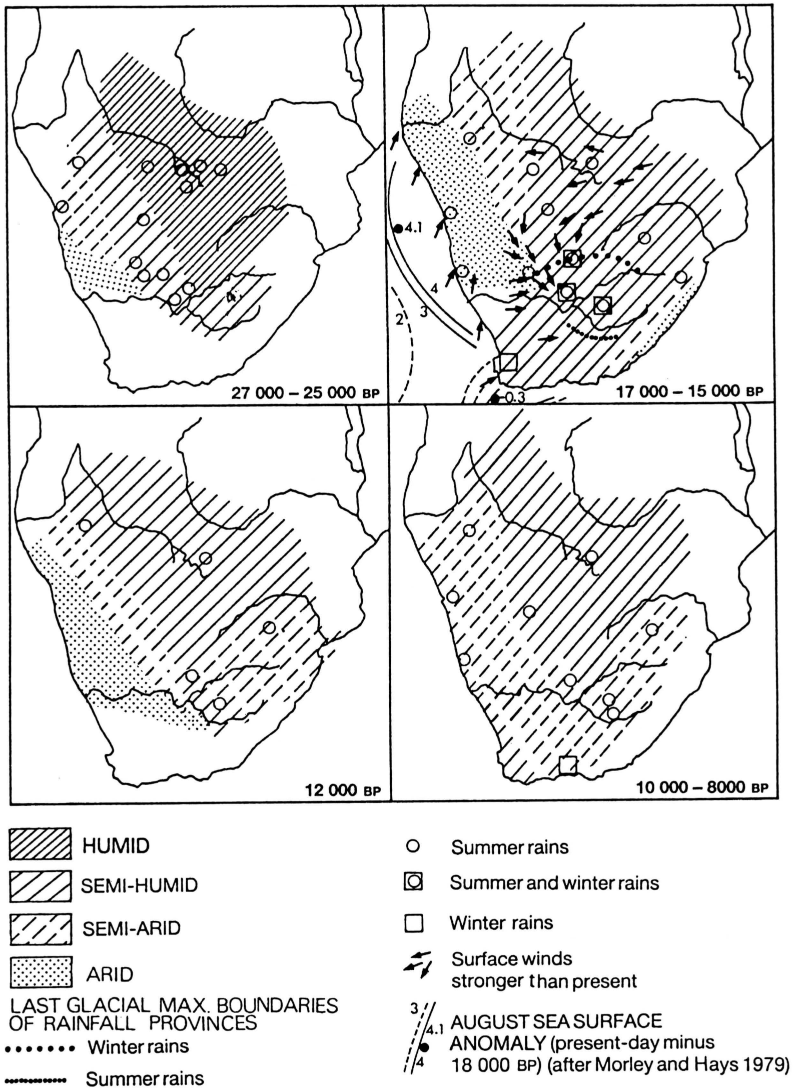

Fig. 2.8 Hypothetical climatic reconstructions for the Kalahari at various times between 27 000 and 8000 years BP, proposed by Heine (1982). Based on radiocarbon dated deposits, molluscs, cave sinters, etc. (Reprinted from: *Palaeoecology of Africa*, (ed. J. C. Vogel, E. A. Voigt, and T. C. Partridge) Volume 15. Southern African Society for Quaternary Research, 1982. A. A. Balkema, P.O. Box 1675, Rotterdam, Netherlands.)

Oscillation could account for the climatic changes observed during the Last Glacial Maximum in Australia and New Zealand, and Salinger (1984), Cockcroft *et al.* (1987), and Tyson (1986) have suggested that a Southern Oscillation pattern could explain Southern African Last Glacial Maximum climates too.

On the basis of present-day synoptic patterns and rainfall variations, they suggest that the summer rainfall region in Southern Africa was drier and the winter rainfall region was wetter during the Last Glacial Maximum. At this time the Southern Oscillation (Low Phase) was enhanced over a period of several thousand years, coinciding with an eastward movement of the Walker circulation over the South Indian Ocean and Malagasy, and along the west coast, a northward movement of the boundary between the winter and summer rainfall regimes from its present position at 30° south to about 25° south (Fig. 2.3). Cockcroft *et al.* (1987) base this latter estimate on the results of Howard and Prell (1984), and Howard (1985) which suggest that the Subtropical Convergence between Africa and Australia shifted between 3 and 5° equatorwards of its present position. This is at variance with the findings of Morley and Hays (1979) who saw a much smaller shift of the Subtropical Convergence in the South Atlantic Ocean during the Last Glacial Maximum. This anomaly may make it necessary to re-evaluate the suggestion that the winter rainfall boundary moved significantly northwards, particularly in view of the absence of good evidence for higher rainfall in that region during the Last Glacial Maximum.

Other factors that have not played a major part in the models discussed above but would have affected weather and climatic patterns of all kinds during glacial times should also be taken into account. For example, Tyson (1977) has suggested that cool periods are generally correlated with enhanced climatic instability and an increase in meridional blocking leading to greater extremes of climate. In Australia, summer temperatures in the Last Glacial Maximum are estimated to have been hotter and winters cooler than the present. Combined with greater wind speeds, this led to drier and generally more extreme weather conditions over much of the continent (Bowler 1976; Ash and Wasson 1983). In addition, several researchers (Gates 1976, p. 1143; Williams 1978, p. 192; Sarnthein 1978; Heath 1979) have pointed out that cooler oceans would have led to lower evaporation rates with a resultant drop in the amount of moisture available for precipitation. Gates (1976) has estimated that this would have been about 23 per cent lower in July in the southern hemisphere and 37 per cent lower in the northern hemisphere 18 000 years ago. A reduction in the amount of moisture evaporated from cooler oceans could have been felt in the winter rainfall region where the winter cyclones could have been both windier and drier. In the all-year rainfall belt on the southern Cape coast a weaker Agulhas current (Martin 1981) could have led to cooler ocean temperatures and the exposed continental shelf would have placed the coastline 80–100 km south of its present position thereby reducing the effect of orographic rainfall that at present accounts for a significant proportion of the annual precipitation along the southern Cape coast.

2.2.3 Test implications for palaeoclimatic models

There is some variability in the reasons for the reconstructions made by the three models, but all agree that Last Glacial climates were probably wetter, windier, and cooler in the winter rainfall region which extended further north and east than at present, and that the summer rainfall region was cooler and probably drier. Where mid-Holocene climates are discussed, the situation is reversed.

Most researchers are in agreement that the concept of an equatorward displacement of circulation belts is too simplistic an explanation, so the question is whether the stronger circulation model or the model that proposes a long-term Southern Oscillation and longitudinal displacement of the Walker circulation is the more appropriate. The implication is that if the former model is correct, then the summer rainfall region would have been moister than if Southern Oscillation drought conditions had pertained. In both models, it is necessary to demonstrate that higher winter rainfall was received north and east of the present-day winter rainfall region, and that the summer and winter rainfall regions were out of phase.

4 Palaeoenvironments in the desert and semi-desert ecozones

4.1 Introduction

In reviewing the evidence for Late Quaternary palaeoenvironments in Southern Africa we have chosen to discuss the data first by rainfall regime (desert, summer, winter, and all-year rainfall regions) and second by ecozone. The ecozone subdivisions have been used by Klein (1980, 1984) and are adopted here because they take account not only of climate, but also of vegetation and fauna, and are therefore more meaningful in palaeoenvironmental reconstructions. The geographic extent of the ecozones is summarized in Fig. 4.1. Radiocarbon dates mentioned in the text are listed in full in Appendix 1.

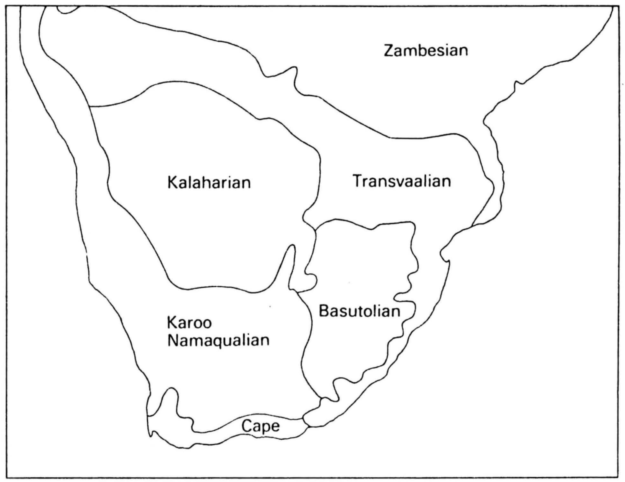

Fig. 4.1 Southern African ecozones (after Klein 1980).

4.2. The Namib

The Namib, with its cool hyperarid to arid climate and unique biota, is sufficiently distinctive to be treated as a separate region within the Karoo–Namaqualian Ecozone (Figs. 4.1 and 4.2). It is the driest region of Southern Africa and has experienced a low amplitude of climatic changes in the Late Pleistocene (Ward *et al.* 1983; Lancaster 1984*b*). It extends for 2000 km along the Atlantic coast of Southern Africa between the Olifants River in South Africa at 32° south to San Nicolau in Angola at 14.30° south. The inland margin of the Namib lies at the base of the Great Escarpment, 100–200 km from the coast. The Benguela Current runs parallel to the coast and strong upwelling of cold sub-Antarctic waters occurs, particularly between latitudes 26 and 28° south.

Benguela

The Namib can be divided into the southern or transitional Namib, a region of rocky plains and low hills with areas of sand sheets and dunes;

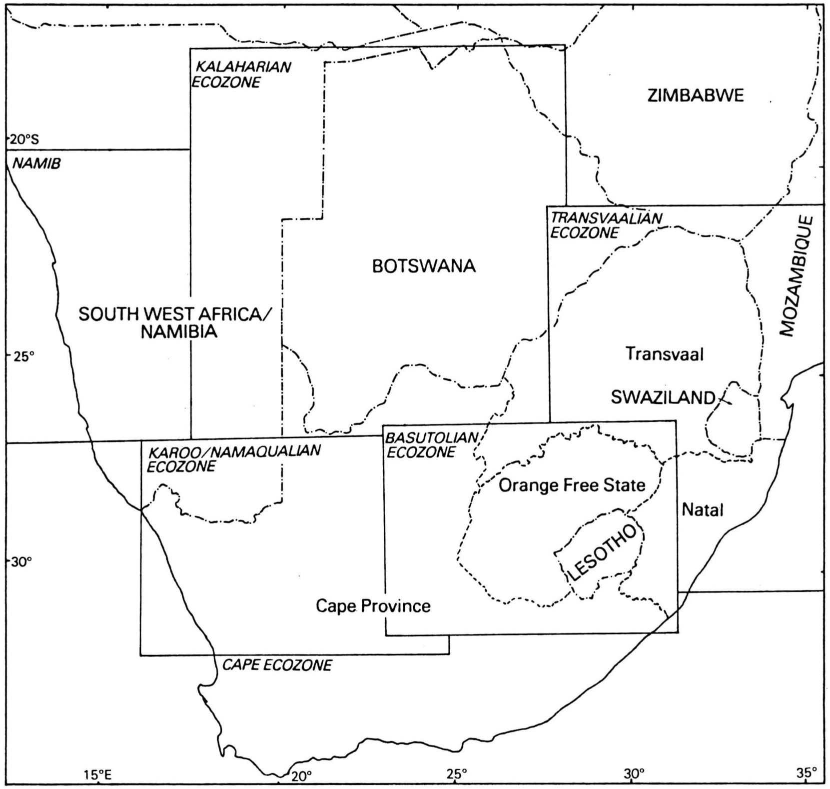

Fig. 4.2 Key to regional ecozone maps in Chapters 4, 5, and 6.

the Namib Sand Sea (Fig. 4.3) between Luderitz and the Kuiseb River; the central Namib plains with scattered ranges of hills and inselbergs which continue north to Damaraland; and the northern Namib where areas of dunes occur along the coast. The climate of the Namib is hyperarid to arid with a steep climatic gradient inland (Seely 1978; Lancaster *et al.* 1984) from the cool foggy coastal zone which receives less than 20 mm of rain annually to the hotter, but less humid areas inland, which receive a scant summer rainfall of 50–100 mm. In the south, the Namib becomes more mesic and grades into the winter rainfall zone of the Cape Province. Throughout the region, vegetation is very sparse. Trees are confined almost entirely to the larger river courses. Grasses spring up rapidly after rains (Seely and Louw 1980), and

climatic gradient

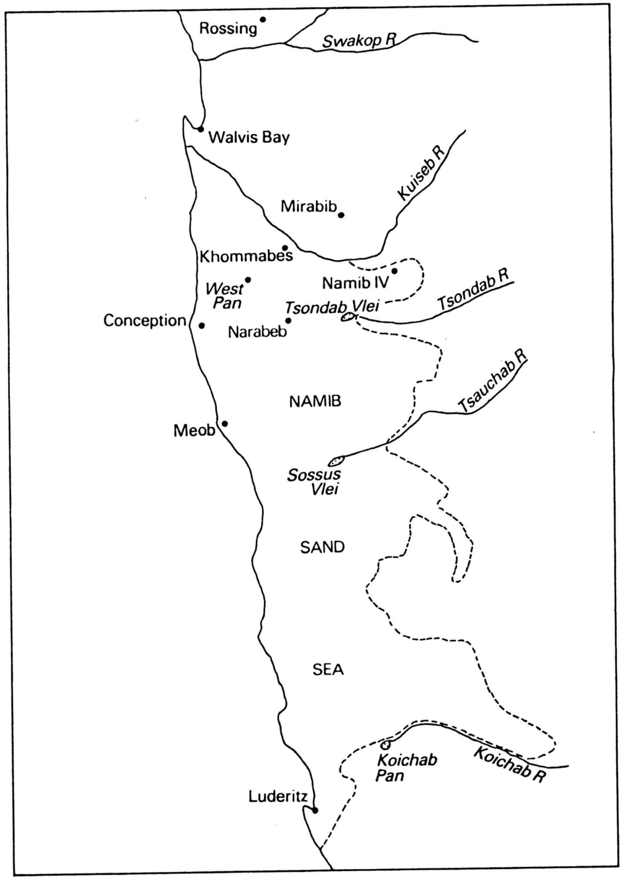

Fig. 4.3 Location of sites in the Namib.

provide grazing for herds of gemsbok (oryx) and springbok, with zebra in the east of the desert.

The causes and age of aridity in the Namib, and their effects upon the evolution of its unique endemic biota have been the subject of considerable debate (Van Zinderen Bakker 1975; Tankard and Rogers 1978; Ward *et al.* 1983). The available geological evidence indicates a long history of arid climates in the region, dating back to the early-mid Tertiary (Van Zinderen Bakker 1975; Ward *et al.* 1983) with hyperarid conditions developed from the late Miocene (7–10 million years ago) onwards (Siesser 1980), following the beginning of strong upwelling by the Benguela Current. Throughout the Quaternary, climatic fluctuations appear to have been of a low amplitude and superimposed upon an arid to hyperarid mean (Lancaster 1984*b*). The surface survival of end-Miocene calcrete palaeosols (Yaalon and Ward 1982) and lacustrine carbonates (Selby *et al.* 1979) suggest that no very humid climates have affected the region in the Late Cainozoic. Ward *et al.* (1983) argue that the Namib has experienced no climate wetter than semi-arid since the end-Miocene, 5 million years ago. Pollen samples from deep sea cores taken on the Walvis Ridge show no significant changes throughout undated Pleistocene sequences, which are dominated by Gramineae, indicating to Caratini and Tissot (1982) that, although there were intervals when the climate was more humid than at present, it was never wet enough to cause a major change in the composition of the vegetation. Similar conclusions were reached by Van Zinderen Bakker (1984) on the basis of a more detailed investigation of Pleistocene pollen from DSDP cores.

Evidence from fluvial deposits (Korn and Martin 1957; Lancaster 1984*b*; Ward 1984*a,b*) and marine sequences (Tankard and Rogers 1978) suggests that there was a progressive increase in aridity throughout the Pleistocene. The average clast size of fluvial deposits in the major river valleys declines from cobbles and gravels in Early Pleistocene deposits to sands and silts in the Late Pleistocene, suggesting a parallel decrease in stream discharge and competence.

In common with many other desert areas, palaeoclimatic evidence is variably preserved in the Namib. There are few depositional basins onshore and preservation of pollen in sediments is poor. As in the Sahara (Rognon 1982) the best preserved evidence comes from the sand seas which show a greater reaction to climatic changes and more contrast in environments between wet and dry periods compared to adjacent gravel plains. As Ward *et al.* (1983) point out, interpretations of palaeoclimates in the Namib must take into account the long narrow shape of the desert, its steep climatic gradient, and the linear oases which penetrate far into the desert along the major river valleys.

4.2.1 Interdune pond and lacustrine deposits

Small areas of mostly thin (<4 m) lacustrine carbonates and associated calcified reeds and gastropod shells are scattered throughout the northern parts of the Namib Sand Sea (Fig. 4.4.) where they are exposed in interdune areas between 50–100 m high linear and star dunes. The deposits, which were mapped as the Khommabes Carbonate Member of the Sossus Sand Formation by Ward (1984*a*), appear to have been laid down at intervals during the accumulation of the Namib Sand Sea.

Sediments of the Khommabes Carbonate Member occur in shallow basins developed in the Tertiary Tsondab Sandstone Formation and overlying Late Tertiary–Early Pleistocene Karpfenkliff conglomerates and gravels of the proto-Kuiseb River (Ward 1984*a*). The pond and lacustrine deposits often display a characteristic association of facies, with white powdery lacustrine carbonates in the centre of the basin, and case hardened deposits and calcite cemented reed stems and organic debris overlying bleached and mottled aeolian dune and interdune sands on the margins (Teller and Lancaster 1986*a*).

Stone Age artefacts of Earlier to Later Stone Age affinities occur in association with the deposits (Shackley 1985) and indicate a long history of episodic and localized ponding of surface runoff and groundwater seepage. The occurrence of artefacts of different ages (especially Earlier

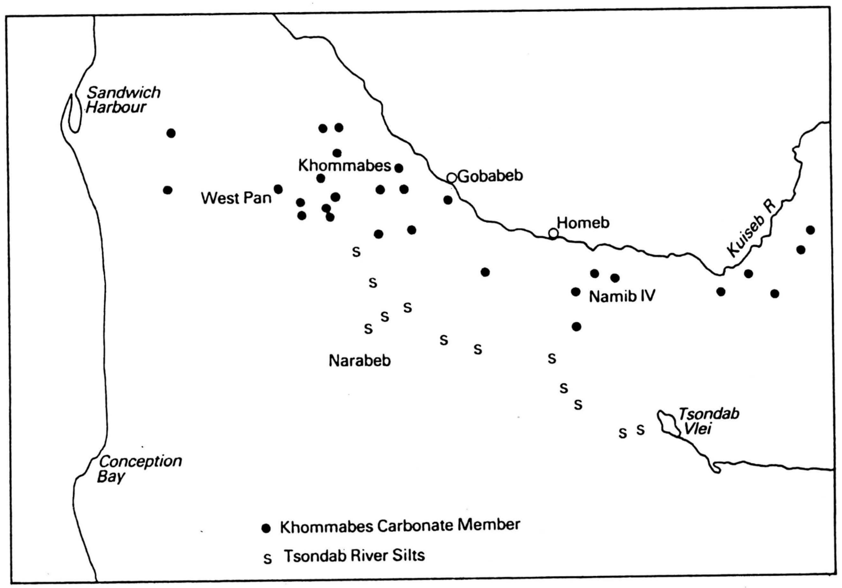

Fig. 4.4 Exposures of interdune pond and marsh deposits in the northern part of the Namib Sand Sea (after Ward 1984*a*).

and Middle Stone Age) at the same site implies reoccupation at intervals when moisture was available within the now arid sand sea. The oldest deposits occur at Namib IV where bones of *Elephas recki* indicate an age of between 700 000 and 400 000 BP (Shackley 1980).

Snails from lacustrine deposits at West Pan include *Melanoides tuberculata, Tomichia ventricosa, Biomphalaria pfeifferei* and *Bulinus* sp. which indicate that fresh to brackish water bodies occurred in interdune areas some 27 500 BP (Teller and Lancaster 1986*a*).

Deposits at Khommabes, a shallow basin situated in an interdune area between 60–100 m high linear dunes 1 km south of the Kuiseb River, were mapped by Ward (1984*a*) and studied in detail by Teller and Lancaster (1985). They record the interaction of fluvial, aeolian, and lacustrine processes at the northern margin of the Namib Sand Sea. The earliest fluvial deposits (Units B and C) consist of sands and silty sands with a high percentage of biotite mica which were deposited in a period of wetter climates by a more extensive flood plain of the Kuiseb River, possibly before the large south–north trending linear dunes has reached this area. They were later cemented by pedogenic calcretes which were subsequently altered to dolocrete. Following this period, the aeolian sands of Unit D, possibly derived by deflation of the flood plain, were deposited on the basin margins. Renewed wetter conditions, or percolation of water from the Kuiseb River, resulted in the growth of reeds and sedges (*Phragmites* and *Juncellus* spp.) as well as nara bushes (*Acanthosicyos horrida*) around the margins of a shallow pond or seepage area and precipitation of calcrete on or in the previous sedimentary sequence as Unit E. Radiocarbon dates from calcified reed stems range from 31 900 to 27 400 BP (Pta-2588, 2589, 2590) and indicate that the wetter conditions prevailed prior to 32 000 BP. In a subsequent period of drier climates, north-easterly dipping cross-bedded aeolian sands (Unit F), similar in composition to the modern dunes, prograded across the carbonates. They provide the first evidence for sand dunes in this location. Radiocarbon dates (Pta-1091, 2584, 2604) from roots, and pedotubules developed in these sands range between 22 400 and 20 900 BP indicating that the sands were probably deposited about this time. Much of the original sedimentary sequence at Khommabes has been deflated and the carbonates of Unit E now form a case-hardened cap over most of the remaining sediments.

Radiocarbon dates from similar pond and reed bed deposits near Gobabeb cluster around 21 500 BP (Pta-2651, 2652) whilst those from Koichab Pan on the southern margins of the sand sea have dates of 23 800–23 300 BP (Pta-3184, 3185). At Kannikwa, in the southern Namib near Port Nolloth, a peat lens from a pan has a radiocarbon date

of 23 200 ± 300 BP (Pta-3978) indicating cool and wet conditions at this time (Beaumont 1986).

Calcified reed beds and organic materials also occur along the Namib coast, adjacent to sites at which vegetation growth today is promoted by seepage of groundwater where the contact between dune sand and the underlying Precambrian bedrock comes to the surface. Calcified organic deposits and *Phragmites* reeds with radiocarbon dates ranging between 12 600±140 and 11 700±120 BP (Pta-1238, 1647, 1830, 1831) occur at Conception and Meob, and indicate that increased groundwater discharge occurred prior to 12 600–11 700 BP, when desiccation and calcification of the reeds took place (Vogel and Visser 1981).

A 36-m thick sequence of calcareous mudstones and interbedded sands at Narabeb, some 40 km south of the Kuiseb River, occurs in a similar interdune situation to deposits of the Khommabes Carbonate Member, but provides a very different record of past climates. The deposits were first described by Seely and Sandelowsky (1974) who suggested that they had been deposited at the former end point of the Tsondab River, which now terminates 40 km to the east at Tsondab Vlei. Selby *et al.* (1979) published uranium-234/thorium-230 dates from a lower member of the deposits, which they regard as having been laid down in a short-lived interdune pond. Six calcareous mudstone units were identified by Teller and Lancaster (1986*b*). They are invariably laminated and sandy, and contain halite filled fractures and desiccation cracks. Unit IV contains planktonic diatoms, mainly *Tabellaria fenestrata*, which are characteristic of eutrophic freshwater lakes. The sands are slightly calcareous, but are otherwise similar to those of the Tsondab Sandstone Formation, which underlies the deposits, as well as to the adjacent dunes. Some sandy units contain chips of calcareous muds deflated from nearby lake beds. Teller and Lancaster (1986*b*) argue that the deposits represent an alternation of wet and dry periods. The calcareous mudstones were deposited in a shallow lake at a former end point of the Tsondab River, which was dammed by dunes. The sands represent aeolian deposition in intervening dry periods. Desiccation is evidenced by mud cracks and halite in the upper parts of the mudstone units. The age of the Narabeb deposits is uncertain. Four radiocarbon dates on the mudstones range betwen 26 000±400 and 20 320±300 BP (Pta-3704, 3759, Beta-9115, 9116), but the uranium-234/thorium-230 dates obtained by Selby *et al.* (1979) from what appears to be the basal mudstone suggest an age of more than 200 000 years for the whole sequence. The association of archaeological materials of Earlier and Middle Stone Age affinities with the deposits (Shackley 1985) indicates that they are more than 25 000 years old.

The present end point of the Tsondab River is Tsondab Vlei, to which the river floods ephemerally. The penetration of the Tsondab River to Narabeb clearly indicates increased flooding and greater discharge as a result of higher rainfall in the highland parts of its catchment (Lancaster 1984*a*) in the period of 30–20 000 years ago, as well as the absence of dunes east of Narabeb. By 14 000 BP its terminus had shifted eastward to a point 5–10 km west of Tsondab Vlei, as evidenced by radiocarbon dates of 14 300±130 and 13 300±90 BP (Pta-1043, 1502) on gastropods from 2–3 m thick silt deposits (Vogel and Visser 1981). Since then, the course of the Tsondab River has been blocked by several 50–100 m high dunes and Holocene sedimentation has been confined to the vicinity of the present vlei (Lancaster 1984*a*).

4.2.2 Fluvial deposits

The Namib Desert is crossed by a number of rivers draining from the highland areas to the east. Apart from the perennial Kunene and Orange rivers, both of which rise well outside the arid zone, most flow ephemerally. From the Kuiseb northwards, most of the large rivers have been known to reach the Atlantic on occasions. Between the Kuiseb and the Orange, no streams reach the sea. This is the result of the increasing aridity of the highland catchment areas and the presence of the dunes of the Namib Sand Sea. The Tsondab, Tsauchab, and Tsams rivers penetrate up to 60 km into the sand sea, to end against the dunes in extensive playas. Most of the Namib rivers have deposits in their valleys which record a sequence of aggradation and incision spanning much of the Late Cainozoic. The importance of the fluvial sequences as indicators of past climates was noted by earlier workers like Gevers (1936), Mabbutt (1952), and Korn and Martin (1957). Subsequent work by Rust and Wienecke (1974, 1980), Marker (1977*a,b*), Ollier (1977), Ward (1982, 1984*a,b*), Lancaster (1984*a*), and Heine (1985) has resulted in the elucidation of the stratigraphy of these sequences, especially for the Kuiseb, but failed to establish whether they reflect climatic or tectonic changes in the Namib and its hinterland. However, it is clear that all the fluvial deposits were laid down in arid to semi-arid climates (Ward *et al.* 1983). Most of the deposits are of early to mid-Pleistocene age (Ward 1984*b*), but in the Kuiseb valley 30 m of micaceous silts (Marker and Muller 1978) accumulated in the period 23 000–19 000 BP (Vogel 1982) in conditions of low stream discharge and competence. The deposits were laid down on a low energy flood plain in a section of the valley upstream from Gobabeb (Ward 1984*b*), rather than at the end point of the river (Vogel 1982) or behind a dune dam (Rust and Wienecke

1974), as evidenced by the abundance of cross-bedded sands of fluvial and derived aeolian origins (Ward 1984*b*). They indicate low stream energy and less intensive floods suggesting drier conditions.

4.2.3 Speleothems from Rössing Mountain

Heine and Geyh (1984) report radiocarbon dates on speleothems from a small limestone cave near Rössing Mountain, 35 km inland from Swakopmund on the hyperarid Namib coast. They occur in an area where rainfall is currently less than 20 mm per year. According to Heine and Geyh (1984) the development of speleothems in this area indicates increased rainfall and run-off in the immediate catchment area of the cave. A series of ten radiocarbon dates indicates that the most recent speleothems were deposited in a period of more humid climates between 41 500 and 26 500 BP.

4.2.4 Micromammals

A sample of micromammalian bones that dates within the last 6000 years from Mirabib rock shelter in the central Namib plains was studied by Brain and Brain (1977). Using the presence of the golden mole (*Eremitalpa granti namibensis*) as an indicator of sandy habitats, it appears that the Kuiseb River, 24 km to the south, has remained a barrier to dune sand during the last 6000 years. The presence of *Malacothrix typica* and a higher relative frequency of gerbils in all but the uppermost units suggest that a more favourable habitat, with moister conditions and more grass cover, was present in the area until recently. Drier conditions than those in the area today have occurred twice in the last 500 years, whilst a period with climates similar to today was centred around 5200 BP.

4.2.5 Stable carbon isotopes

In contrast to the changes observed in the carbon-13 content of bone collagen in zebra teeth from Melikane Cave in Lesotho, zebra teeth from Late Pleistocene (*c.* 70 000–20 000 BP) and Holocene (7000 BP) units at Apollo 11 Cave excavated by Wendt (1972) on the eastern margin of the southern Namib Desert show no appreciable difference (Table 4.1). This led Vogel (1983) to suggest that C_3 grasses did not spread into southern Namibia during the Late Pleistocene, indicating that winter rainfall did not extend much further north in the Late Pleistocene than it does today.

Table 4.1. Carbon-13 content of collagen from *Equus* teeth excavated from Apollo 11 Cave in southern South-West Africa/Namibia by Wendt (1972). The last column lists the contribution of C_3 (winter rainfall) plants to the diet of the animals as deduced from the isotopic composition of the teeth. The results show no appreciable difference between the Holocene and Late Pleistocene samples (from Vogel 1983).

Sample	Approx age (years BP)	Carbon-13 content	% C_3 in diet
B156/Z1	Recent	−12.6	36
B157/Z2	7000	−14.1	47
B158/Z3	20 000	−14.9	53
B159/Z4	70 000	−14.8	53

4.3 Chronological summary

Presently available dated palaeoclimatic information for this ecozone is confined to the last 32 000 years. In the period from prior to 32 000 to 20 000 BP the region appears to have been substantially wetter than today with interdune lake and pond deposits in the Namib Sand Sea. River flow was also stronger than at present. A drying trend was present locally after 27 000–26 000 BP. Cool conditions are evident around 23 000 BP. The period spanning the Last Glacial Maximum (20 000–16 000 BP) appears to have been dry in the Namib. The interval from 16 000 to 12 000 BP sees a renewal of increased moisture availability in the Namib, continuing until around 11 000 BP. For the Holocene, the climate appears to have been slightly moister than it is today, with a dry interval around 2500 BP.

4.4 The Karoo and Bushmanland

The Karoo and Bushmanland also form part of the Karoo–Namaqualian Ecozone. Until very recently, virtually no palaeoclimatic evidence was recorded from this part of Southern Africa, although the widespread occurrence of diatomite deposits in pans was noted by Kent (1947). Recent investigations of pan deposits and archaeological surveys have indicated that a wealth of palaeoclimatic evidence exists in this region and may prove valuable in the correlation of sequences and events in the summer and winter rainfall areas of the subcontinent.

7 Summary and conclusions

Reliable information on the climatic history of Southern Africa over the last 130 000 years has not been easily won and data collection has inevitably been influenced by prevailing hypotheses. Models based on the European glacial sequence and the pluvial hypothesis sought data on past rainfall changes from the sedimentological and geomorphological record. By the time this model was dropped in favour of one based on the equatorward movement of climatic belts during glacial cycles, biological data were available, and both temperature and rainfall changes could be traced. In the past few decades, deep sea cores have given more detail on global sea temperatures, but in Southern Africa this record is in need of systematic testing in land-based sequences. Isotope records in speleothems and aquifers are being used for this purpose and information is being sought on the response of geomorphic, sedimentary, and biological systems to the periodic changes in climate during the Quaternary. The results summarized here give an idea of how much, and how little, we know.

The radiocarbon dates that relate to palaeoclimatic data described in this book are summarized in Appendix 1. Figures 7.1, 7.2, and 7.3 use radiocarbon dated occurrences that have given reliable *primary* palaeoclimatic data as defined by Hecht *et al.* (1979) (see chapter 1.1). In the summary that follows, the evidence for climatic change during the past 130 000 years is discussed on an inter-regional basis. The time periods are loosely linked to the oxygen isotope chronology, but do not imply a formal correlation.

7.1 *c.* 130 000–80 000 BP

The most securely dated site in this time range is the Cave 1/1a complex at Klasies River Mouth on the southern Cape coast (Singer and Wymer 1982). The Middle Stone Age (MSA) deposits have accumulated on the floors of Caves 1 and 1b at 6–8 m above present sea level. The caves were formed by several Pleistocene high sea levels and the pebbly beach at the back of Cave 1 probably relates to one of these (Deacon *et al.*

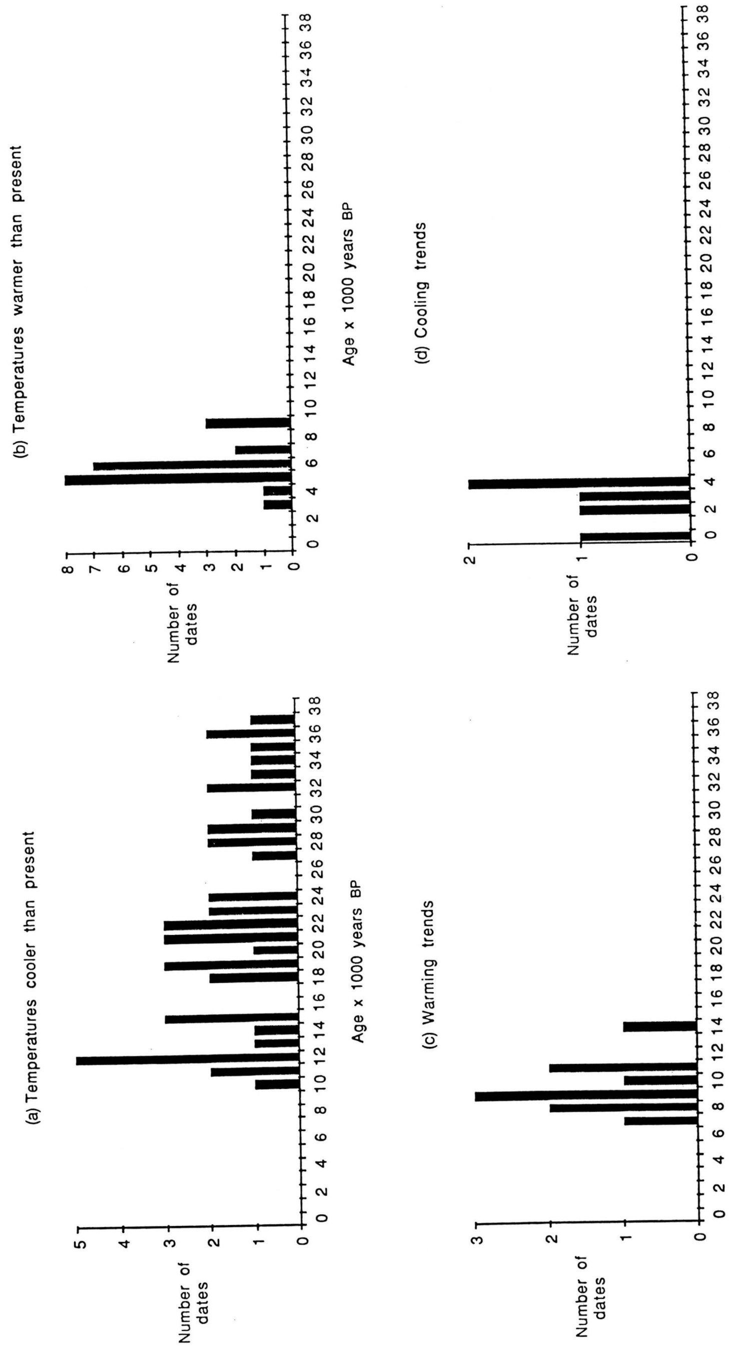

Fig. 7.1 Radiocarbon dated instances of past temperature conditions in Southern Africa (a) cooler than at present, (b) warmer than at present, (c) showing a warming trend, i.e. conditions warmer than the previous period, but not as warm as at present, and (d) showing a cooling trend, i.e. cooler than the previous period.

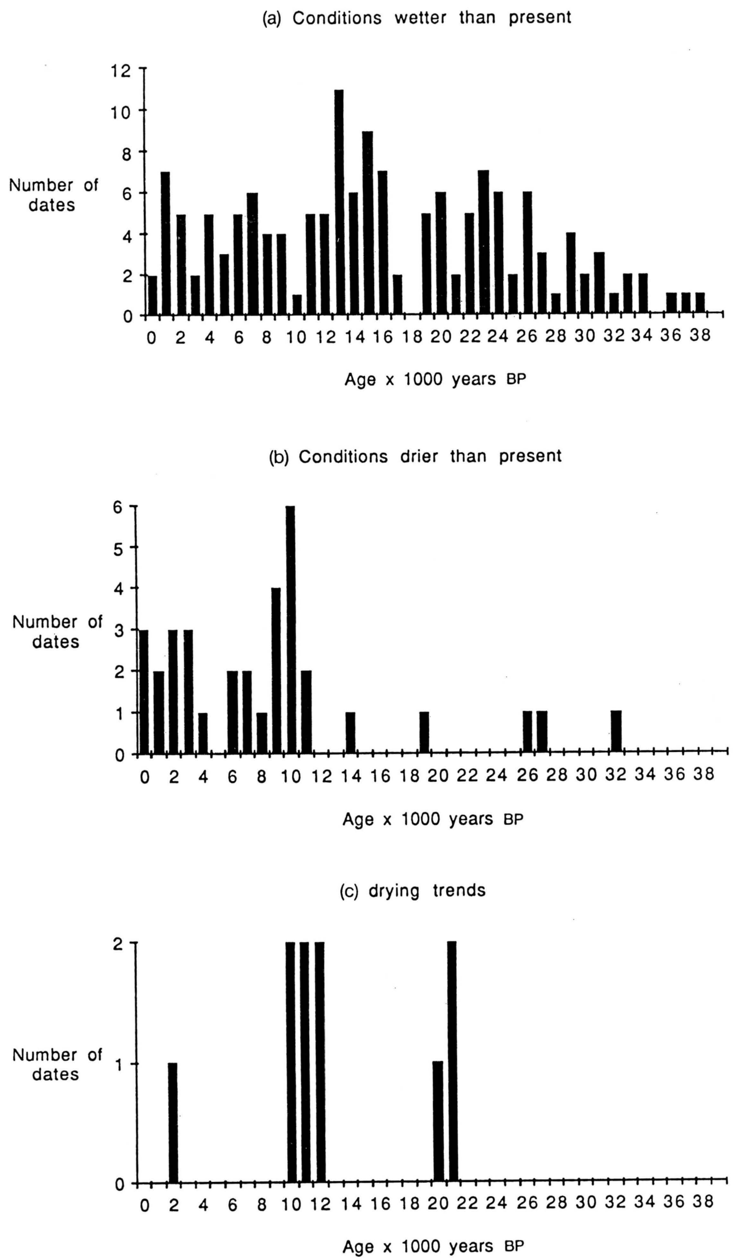

Fig. 7.2 Radiocarbon dated instances of past climatic conditions in the summer rainfall zone (a) wetter than at present, (b) drier than at present, and (c) showing a drying trend.

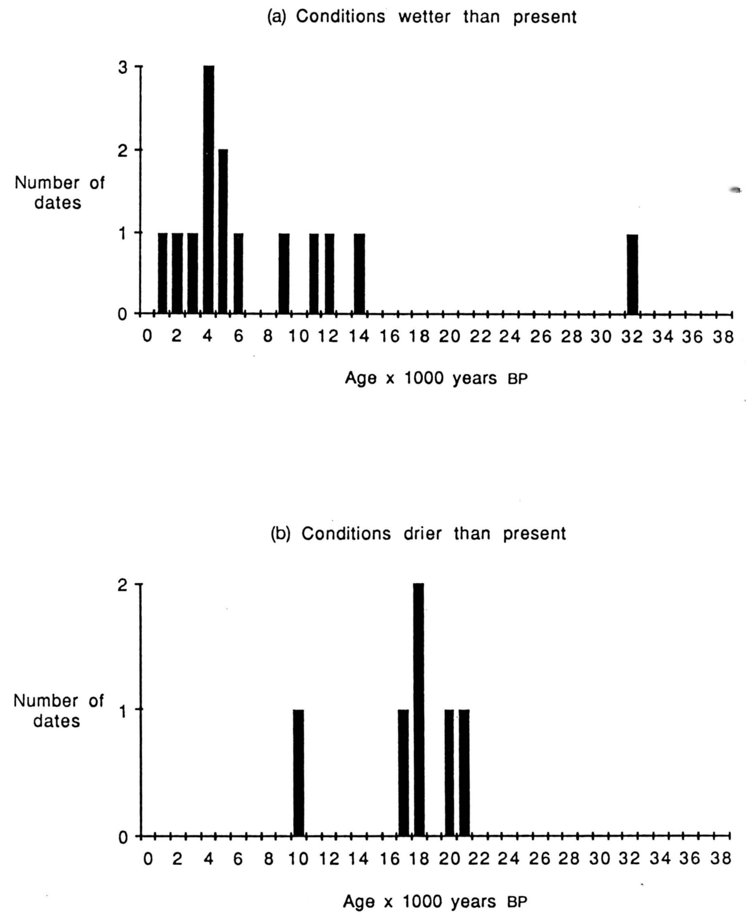

Fig. 7.3 Radiocarbon dated instances of past climatic conditions in the winter and all-year rainfall zone (a) wetter than at present, and (b) drier than at present.

1986). The large mammals and shellfish species in the lowermost (MSA I and MSA II) deposits are similar to those of the Present Interglacial suggesting that they either coincide with or immediately post-date the warmest phase of the Last Interglacial. In the overlying units (Howiesons Poort, and MSA III and IV) the fauna is dominated by grazers rather than browsers which have been shown to characterize most of the Last Glacial cycle in the Cape Ecozone (Klein 1980, 1983) and indicate cooler temperatures which promoted grasslands rather than the bushier environments typical of interglacials. The oxygen isotope ratios of marine shells (Shackleton 1982; H. J. Deacon, personal communication) show initially warm temperatures in the MSA I and II with an isotopic signature typical of the Last Interglacial (stage 5e) at the base and cooler, but not cold temperatures in the overlying samples. Sub-stages within stage 5 are recognizable in the cave sediments, in the marine shells and in the micromammalian taxa. The recent excavations (Deacon *et al.* 1986) give a more detailed picture of changes in the marine shell and micromammalian samples that show a major regression beginning at the top of the MSA II units and reaching a peak at the top of the Howiesons Poort. With uranium disequilibrium and aspartic acid dates confirming that this part of the sequence is younger than 100 000 years, the regression has been correlated with the Huon Peninsula sea level curve (Chappell and Shackleton 1986) dating it to between 80 000 and 60 000 BP at the end of the Last Interglacial (H. J. Deacon, personal communication).

At inland sites such as Border Cave, Alexandersfontein, Gaap Escarpment, Rose Cottage, and the Lower Vaal River terraces, there are either no faunal remains and/or no formations such as raised beaches to tie the observations directly into the oxygen isotope sequence. Instead, it is the alternation of changes in tufa formation, sediments, and the species frequencies of small and larger mammals, interpreted as the result of gross temperature fluctuations, that is assumed to represent the responses to the warmer and cooler sub-stages of oxygen isotope stage 5. Unlike the Cape Ecozone, an increase in browsers and micromammalian taxa that prefer bushy conditions at Border Cave is taken to indicate cooler rather than warmer temperatures (Klein 1977). In the case of other sites, increased run-off in the Riverton Formation and tufa formation on the Gaap Escarpment, which represent changes in the rainfall regime, are assumed to relate to stage 5 and its subdivisions because of their association with, or position in relation to, Middle Stone Age occupation deposits (Butzer 1984*a,b*). The dating of wetter periods on the Gaap Escarpment is more secure with periods of tufa formation dated by thorium to 103 000 ± 6000 BP and by uranium to 86 000 ± 21 000 BP (Vogel and Partridge 1984, p. 512), but the standard errors are large. There are no well dated deposits with palaeoenvironmental data from the winter rainfall region to compare with those to the north and east.

7.2 *c.* 80 000–50 000 BP

This time period is still beyond the resolution of radiocarbon dating and the assignment of deposits to it depends very largely on their stratigraphic position in long sequences or on the associated stone artefacts.

The most complete sequence comes from Boomplaas Cave in the southern Cape with additional data from Die Kelders and Klasies River in the Cape Ecozone and Border Cave in the Transvaalian Ecozone. At all three, biological data and cave sediments suggest cooler conditions than the preceding time period. The faunal remains from the Cape sites all indicate grassland rather than bushy vegetation, and at Boomplaas the relatively high incidence of *Pelea capreolus*, the vaalribbok, indicates cool temperatures as does the range of micromammalian species (Klein 1976, 1978*b*; Avery 1982*a,b*; Deacon *et al.* 1984*a*). Temperatures derived from factor scores on Boomplaas microfauna suggest at least one warmer peak within cool conditions in this general period (Thackeray, in press). At Border Cave the fauna is represented by taxa which prefer bushier environments (Klein 1977; Beaumont *et al.* 1978; Avery 1982*b*), and at Redcliff in Zimbabwe cooler temperatures are indicated by the extension of the range of some larger mammals which only occur further south today (Cruz-Uribe 1983). Cold intervals when temperatures were low enough to induce frost spall horizons in cave sediments may be dated to this time range at Melikane, Rose Cottage, Bushman Rock Shelter, Border Cave, and several sites in the southern Cape, for example Nelson Bay Cave, Boomplaas, Diepkloof, Die Kelders and Klasies River (Carter 1976; Butzer 1984*a,b*; Deacon *et al.* 1984).

Moister conditions are apparently indicated by tufa formation on the Gaap Escarpment dated by thorium to 61 000 ± 5000 BP (Vogel and Partridge 1984, p. 512). Undated charcoal samples from this time range at Boomplaas and micromammals at Die Kelders both suggest it was drier (Avery 1982*a*; Scholtz 1986) and the Boomplaas wood anatomy is suggestive of windy conditions as well (Scholtz 1986).

7.3 *c.* 50 000–24 000 BP

The oxygen isotope record in both the Wolkberg and Cango caves stalagmites begins within this time period. Wolkberg, with fewer readings, shows a gradual cooling from about 40 000 to about 20 000 BP (Talma *et al.* 1974). Cango shows a slight cooling trend from *c.* 45 000 to 30 000, and then a cool but fairly stable record up to about 22 000 BP (Vogel 1983). The carbon-13 content of the Cango stalagmite also confirms cooler temperatures over this time range with C_3 plants predominating (Vogel and Talma, personal communication). The fossil water from the Uitenhage aquifer dates from about 28 000 BP and also shows fairly stable temperatures from the beginning of the sequence up to about 20 000 BP (Heaton *et al.* 1986*a*). At Boomplaas in the charcoal and micromammal data there is the suggestion of a slightly warmer interval centered around 32 000 BP (Figs 6.5 and 6.7) (Deacon *et al.* 1984*a*) that is possibly present also in the carbon-13 record from Cango Caves (Vogel and Talma, personal communication). By contrast, frost spalling is reported from Border Cave between 38 000 and 33 000 BP, from Wonderwerk Cave between *c.* 30 000 and 26 000 BP, and at several cave sites in Lesotho, including Melikane (Carter 1976; Butzer *et al.* 1978*a*; Butzer 1984*b*). Mechanical weathering on the Gaap Escarpment before the formation of Tufa II is interpreted by Butzer (1984*b*) as indicative of increased frost and moister conditions around 30 000 BP.

In several ecozones this period is characterized by the wettest conditions of the past 130 000 years. Maximum lake levels were reached in the Kalaharian and Karoo-Namaqualian Ecozones. The Makgadikgadi lake levels are reported to have been markedly higher at various intervals between 40 000 and 20 000 BP (Cooke 1980, 1984; Heine 1978, 1982). Higher water tables in interdune areas in the Namib and stabilized dunes in the south-western Kalahari, both indicating much moister conditions, are dated to between about 32 000 and 28 000 BP (Heine 1982; Teller and Lancaster 1985). Cave sinters at Rössing between *c.* 41 000 and 26 000 BP also suggest increased humidity (Heine 1984; Heine and Geyh 1984). In the Bushmanland area of the Karoo-Namaqualian Ecozone, shell deposits have been interpreted as evidence for higher water levels at Breek Been Kolk between *c.* 38 000 and 14 000 BP (Kent and Gribnitz 1985; Beaumont 1986). At Wonderkrater in the Transvaalian Ecozone higher precipitation is indicated by the expansion of woodland and forest (Scott 1982*a*).

In the Cape Ecozone in the all-year rainfall region, more effective precipitation is suggested by the charcoals from Boomplaas Cave which include a larger number of woodland taxa (Deacon *et al.* 1984) and by the high incidence of *Myosorex varius* amongst the micromammals (Avery 1982*a*, 1983). Xylem analysis of the charcoals suggests it was cold with relatively high winter rainfall (Scholtz 1986).

Northern hemisphere palaeoenvironmental data summarized by Kukla *et al.* (1981) and oxygen isotope data from Indian Ocean deep sea cores (Chappell and Shackleton 1986) suggest that we may expect land-based sites dating to this time period to have been cooler than in the preceding 30 000 years, but there is little clear evidence for this in Southern Africa. Widespread evidence for moister conditions argues against severely cold temperatures, and while it was still cool enough to sustain grassland rather than browse in the southern Cape, and more C_3 plants grew in the highlands of Lesotho and the southern Cape than occur there at present (Vogel 1983, personal communication), the larger mammal fauna in parts of the interior (particularly around the eastern Transvaal at Heuningneskrans) was essentially similar to that of today. The same range of larger mammals occurs in the human occupation layers of that time as occurred during the Holocene levels at the same sites, but this may merely emphasize the fact that the larger mammals in the eastern part of the summer rainfall region are not a particularly sensitive indicator of past climatic changes.

7.4 *c.* 24 000–16 000 BP

Without exception, all palaeoclimatic data relating to this time period concur that it was the coldest interval of the last 130 000 years. Estimates of the temperature reduction vary from 8–9.5°C at Wolkberg in the Transvaal (Talma *et al.* 1974) to 5.5 and 5°C lower than at present in the Uitenhage aquifer and the Cango Caves speleothem respectively (Vogel 1983; Heaton *et al.* 1986*a*).

For the most part conditions were also drier, as predicted in general circulation models. In the Namib and Kalahari evidence points to falling lake and groundwater levels culminating in dry conditions around 19 000–18 000 BP (Heine 1978; Helgren and Brooks 1983; Teller and Lancaster 1985). The end of the long-continued sequence of high lake levels is marked by a relatively large number of radiocarbon dates for this event and should be interpreted as an indication of the onset of drier conditions rather than a marked wet period. Drier conditions are reported also from Haaskraal in the Karoo (Partridge and Dalbey 1986) and Kathu Pan in the southern Kalaharian Ecozone (Beaumont *et al.* 1984). To the east in the Basutolian Ecozone, pollen analyses suggest cooler and drier conditions at Cornelia and Craigrossie between *c.* 20 000 and 12 000 BP (Scott 1986). In the winter rainfall region of the western Cape, cool, dry, and windy conditions are evidenced by sediments in Elands Bay Cave (Butzer 1979) and by the size of molerats dating to some time between *c.* 20 000 and 11 000 BP (Klein 1984, 1986).

There are several instances of higher precipitation: from Alexandersfontein in the northern Cape (Butzer 1984*a,b*), from the Blue Pool tufa on the Gaap Escarpment (Butzer *et al.* 1978*b*; Butzer 1984*a,b*) and from Breek Been Kolk, Hoezar Oost and Kannikwa in the Karoo-Namaqualian Ecozone (Beaumont 1986).

Although the evidence from these sites seems to support the case for wetter conditions during the Last Glacial Maximum north and east of the present-day winter rainfall region, the palaeosols that alternate with dunes in the south-western Cape (Butzer 1984*b*) are not considered to be reliable evidence for higher rainfall at this time. They are dated on the basis of one radiocarbon determination on bone from the A horizon in a palaeosol at Melkbosstrand that Butzer (1984*b*, p. 239) considers indicative of surface stability and a better ground cover rather than a genuinely moist climate. Similarly, the radiocarbon dates of *c.* 18 600 and 18 100 BP on charcoal and ostrich eggshell from the ferruginized horizon at Nelson Bay Cave on the southern Cape coast date the human occupation of the site and not the later period of ferruginization. The lack of change in the incidence of C_3 grasses in the diet of zebras from

Apollo 11 cave at 28° south, which shows that the winter rainfall belt did not shift this far northwards (Vogel 1983), is further evidence against the assumption (Cockcroft *et al.* 1987; Tyson 1986) that winter rainfall was intensified and extended as far as 25° south during the Last Glacial Maximum.

The biological data are not only consistent in indicating a cold and generally dry glacial climate, but also show unusually low species diversity in both plant and animal remains which is a consistent feature of harsh Last Glacial Maximum environments. Again the Boomplaas site with complementary evidence from larger mammals, micromammals, charcoal, and pollen, and its relatively complete sequence, provides the key evidence for the Cape Ecozone (Deacon *et al.* 1984*a*). Together the results demonstrate that an overall temperature reduction of the order of 5°C lower than at present, combined with reduced effective precipitation, is sufficient to impoverish both the vegetation and animal communities of the Cape Ecozone to the extent that they have no modern analogue. Similar impoverishment in vegetation at the Last Glacial Maximum has been noted in the Transvaal at Wonderkrater (Scott 1982*a*), but there are insufficient data from other ecozones to indicate whether this phenomenon was widespread in Southern Africa. More generally, the archaeological proxy data, in highlighting the low incidence of occupation sites of Stone Age people in this time range, confirms that hunter-gatherers in most areas of Southern Africa apparently suffered a marked decrease in population numbers during the Last Glacial (Deacon and Thackeray 1984). The north-western Cape seems to be an exception to this pattern (Beaumont 1986).

7.5 *c.* 16 000–10 000 BP

There is evidence from a wide range of environments for markedly wetter conditions immediately after the Last Glacial Maximum between *c.* 16 000 and 11 000 BP. In the Kalahari and Namib it was wetter at Conception, Meob, Otjimaruru, Gobabis, Urwi Pan, Drotsky's Cave, and the Okavango, but conditions were not as wet as they were between 40 000 and 20 000 BP (Lancaster 1979; Vogel and Visser 1981; Butzer 1984*a,b*; Shaw 1986). In the southern Cape at Boomplaas Cave, on the other hand, the period from *c.* 16 000 to 12 000 was the wettest time of the past 70 000 years. From *c.* 12 000 to 8000 BP conditions around Boomplaas became drier with summer drought (Scholtz 1986). Cave speleothem formation in the Cango caves and in Boomplaas Cave ceased at about 16 000 BP so that no estimate of temperature is available between this time and the middle Holocene (Vogel 1983; Deacon *et al.* 1984*a*). Wetter conditions with summer and winter rainfall increases are indicated by arboreal and shrub pollen at Craigrossie in the Basutolian Ecozone between *c.* 12 600 and 10 700 BP (Scott 1986) and cooler and moister conditions are indicated at Aliwal North between *c.* 13 000 and 12 000 BP (Coetzee 1967) and pre-10 000 BP in East Griqualand (Tusenius 1986). In the winter rainfall region, molerat sizes at Elands Bay cave suggest it was wetter between *c.* 11 000 and 9600 BP, but it was drier before that time (Klein 1984, 1986).

There is clearly some variability for there are data in the interior suggesting a short dry phase between moister ones *c.* 14 000 to 13 000 BP at Alexandersfontein in the Basutolian Ecozone (Butzer *et al.* 1973*a*), and on the Gaap Escarpment in the Kalaharian Ecozone between *c.* 13 500 and 11 500 BP (Butzer *et al.* 1978*b*). Drier conditions persisted at Kathu Pan.

The speed of environmental change can be gauged from the speed with which sea levels rose as glaciers melted at higher latitudes after 16 000 BP in the southern hemisphere. Figure 6.4 illustrates the dated sea levels off the Southern African coast compared against the curve drawn from the Huon Peninsula sea levels, New Guinea, that Chappell and Shackleton (1986) have recalculated after detailed correlation with the oxygen isotope record from deep sea core V19-30. The results are reasonably consistent. Dated land-based sequences in Southern Africa (for example, Boomplaas Cave) confirm that the initial shift in temperatures took place between 16 000 and 14 000 BP, and that this shift was coincident with that observed in the Antarctic (Lorius *et al.* 1979, 1985) and other southern hemisphere continents, and preceded that in the northern hemisphere by between 2000 and 3000 years (Salinger 1984, p. 145; Labeyrie *et al.* 1986, p. 705).

Biological data indicate that marked changes in vegetation and small mammals occurred between 16 000 and 14 000 years ago, but that larger mammal changes in the Cape Ecozone were somewhat slower and post-date 12 000 BP when several of the Late Pleistocene 'giant' bovids became extinct, apparently as the result of the combined effects of environmental changes and more efficient equipment of Later Stone Age hunters (Klein 1980, 1984; Avery 1982*a*; Deacon *et al.* 1984*a*). There is little or no evidence for a comparable change in the larger mammal fauna in most other ecozones where essentially the same taxa were present during the Late Pleistocene and the Holocene.

7.6 *c.* 10 000–0 BP

During the Holocene fluctuations in both temperature and humidity occurred, but the scale of change was considerably lower than that observed between the glacial and interglacial modes of the Late Pleistocene and Holocene. Terms such as 'wetter', 'drier', 'warmer', and 'cooler' should therefore be seen only as relative estimates in the context of the Holocene. Some measure of the scale of change involved can be seen in the temperature estimates calculated from the oxygen isotope content of the Cango speleothem after *c.* 6000 BP (Fig. 6.12) (Vogel and Talma, personal communication) and temperatures calculated from factor analysis of micromammalian samples (Fig. 6.7; Thackeray 1987). They both show that temperature fluctuated less than 1°C around the present day mean, although the Wolkberg speleothem from the Transvaal suggests the variability may have been up to ± 2°C (Talma *et al.* 1974).

7.6.1 *c.* 10 000–7000 BP

Whereas some sites with long and detailed sequences show variability in this time span with evidence for slightly cooler and warmer intervals (Aliwal North, Wonderkrater, Byneskranskop, Boomplaas), the overall picture is one of increasingly warm temperatures.

In the Cape, Transvaalian, Basutolian, and Kalaharian ecozones, where precipitation estimates have been made they indicate a generally drier trend, with the exception of moister conditions indicated by charcoals at Siphiso in Swaziland (Prior and Price Williams 1985) and by a seasonally high water table at Rose Cottage Cave (Butzer 1984*a*). In the Kalaharian Ecozone there is evidence for locally active springs at Etosha between *c.* 11 000 and 9000 BP, and at Stampriet between *c.* 14 000 and 8000 BP (Butzer *et al.* 1978*b*; Heine 1982, 1984). There is no good evidence for markedly wetter conditions raising lake levels to the same extent as has been observed in East Africa in the Early Holocene (Nicholson and Flohn 1980).

In the Cape Ecozone (the only ecozone to show marked differences between glacial and interglacial faunas), larger mammal communities were changing gradually and by 5000 BP were essentially similar to those of historic times. Whereas equids and alcelaphines are still found in small numbers in deposits dating to 10 000 BP (Boomplaas, Nelson Bay Cave, Byneskranskop, Melkhoutboom, Wilton), and vaalribbok, mountain reedbuck, and roan antelope were relatively common in the early Holocene, they are largely absent thereafter indicating that the scrub and bush element in the vegetation increased gradually at the expense of grassland (Klein 1980, 1984). In East Griqualand at Colwinton and Ravenscraig temperatures were warm enough after 10 000 BP to allow alcelaphines and equids to spread to grassland above the escarpment whereas they were uncommon there in the Late Holocene (Opperman

1982, 1984). Charcoal from these sites also shows dry conditions prevailed in the Early Holocene (Tusenius 1986).

To the north and east in the summer rainfall region, pollen analyses from Wonderkrater, Scot, Rietvlei, and Moreletta (Scott 1984*a,b*) indicate relatively dry and cool conditions between *c.* 10 000 and 8500 and somewhat warmer temperatures between *c.* 8500 and 7500 BP.

7.6.2 *c.* 7000–4000 BP

There are few continuous sequences in the first half of the Holocene from which to gauge the scale and timing of temperature changes, but what is available indicates this was the warmest time at Wonderkrater (Scott 1982*a*), Kathu Pan, Equus Cave, and Wonderwerk (Beaumont 1986) in the summer rainfall region, and Groenvlei (Martin 1968) and Boomplaas (Deacon *et al.* 1984*a*) in the all-year rainfall region. There is a break in occupation at Boomplaas between *c.* 9000 and 6400 BP, but the temperature estimates from the micromammals (Thackeray 1987; Fig. 6.7) and the charcoal analyses (Scholtz 1986) show the sample from 6400 BP to have been the warmest in the entire sequence with hot and dry summers suggested by the wood anatomy of the charcoal. The Cango speleothem has no dated samples between *c.* 16 000 and *c.* 5500 BP and although there is a peak in the oxygen isotope 'temperature' curve in the Holocene sequence dating to *c.* 2000 BP, this may not be a true reflection of the scale of temperature change (Vogel and Talma, personal communication). The 'temperatures' derived from Wolkberg Cave (Talma *et al.* 1974) and the Uitenhage aquifer (Heaton *et al.* 1986*a*) show a similar peak dating in the latter half of the Holocene.

From the Transvaalian, Basutolian, and Kalaharian ecozones (Wonderkrater, Alexandersfontein, Wonderwerk, Kathu Pan, Equus Cave, and Okavango) there is some indication of increased moisture between about 7500 and 5000 BP. In the southern Cape moister conditions evident from pollen analyses post-date 7000 BP (Martin 1968), and the stabilization of dunes seen in the Beacon Island soil and interpreted as indicating more humid conditions also probably post-dates 8000–7000 BP (Helgren and Butzer 1977). The warm and moist environments of the mid- to Late Holocene in the southern Cape were marked by an increase in forest taxa in pollen spectra from Groenvlei and Norga. This increase can be related tentatively in the mid-Holocene to a rise in sea level of the order of +1.5 m coincident with warmer temperatures evident from associated warm water diatoms at Groenvlei (Martin 1968) and oyster beds in Langebaan Lagoon (Flemming 1977). Shell at Elands Bay (Yates *et al.* 1986) in a cobble beach 2.8 m above present sea level is dated to nearly 4000 BP.

Micromammals from Byneskranskop in the south-western Cape suggest it was relatively dry between *c.* 6400 and 3900 BP (Avery 1982*a*, p. 338), and drier conditions are reported also from the Basutolian Ecozone at Cornelia, Craigrossie (Scott 1986), Bonawe, Colwinton and Ravenscraig (Tusenius 1986), and from the Transvaalian Ecozone at Siphiso in Swaziland (Prior and Price-Williams 1985) and the Kalaharian Ecozone on the Gaap Escarpment (Butzer *et al.* 1978*b*).

7.6.3 *c.* 4000–0 BP

The last 4000 years have seen low level temperature fluctuations around the present-day mean at most sites after 2000 BP (Vogel 1983; Heaton *et al.* 1986). Between 4000 and 3000 BP wetter intervals have been recorded in lake levels and alluvial deposits in the Kalaharian Ecozone at Makgadikgadi (Helgren 1984), Drotsky's Cave (Grey and Cooke 1977), in the Basutolian Ecozone at Alexandersfontein (Butzer *et al.* 1978*b*) and at Cornelia and Craigrossie (Scott 1986) and possibly Rose Cottage (Butzer 1984*a*). In the Transvaalian Ecozone at Wonderkrater (Scott 1982*a*), Border Cave (Avery 1982*a*), Voigtspost (Horowitz *et al.* 1978) and Moreletta (Scott 1984*b*) it seems to have been somewhat drier. These findings are at variance with those from Siphiso, also in the Transvaalian Ecozone, where the charcoal frequencies are interpreted as indicating relatively moist conditions (Prior and Price-Williams 1985). Charcoals from the East Griqualand sites also suggest relatively moist conditions after 3000 BP, and those from Boomplaas suggest warm and mesic conditions with a lower incidence of drought than occurred in Early and mid-Holocene times (Scholtz 1986).

In the Cape Ecozone somewhat cooler conditions than in the mid-Holocene are suggested by the lower incidence of forest taxa and a higher grass component in pollen spectra in the last 2000 years (Martin 1968; Scholtz 1986), whilst micromammals at Byneskranskop and Elands Bay Cave in the south-western and western Cape suggest it was somewhat moister after 2000 and 3000 BP, respectively, and cooler at Die Kelders during the last 2000 years (Avery 1982*a*, 1983). Optimum conditions for forest expansion in the southern Cape occurred between about 6000 and 2000 BP when it was relatively warm and moist, because thereafter forest taxa were never as common again.

The most important feature of the last 5000–4000 years is the fact that where biological data are available, they show that it was only during this time that modern community alliances were formed. The biogeographic implications of this observation are discussed more fully in the next chapter.

7.7 Summary of radiocarbon dated primary evidence

The main features of the most securely dated primary palaeoclimatic data (equivalent to the level 1 data of Hecht *et al.* 1979) are summarized in Figs 7.1, 7.2, and 7.3. The information is, of course, limited to the last 40 000 years when radiocarbon dates are the most reliable. Caution should be used in the interpretation of these data because the number of dated occurrences is still small and they are geographically scattered. Full details of the dataset are given in Appendix 1.

Figure 7.1a shows that conditions were generally cooler than today prior to 11 000–10 000 BP, but were warmer than today 6000–5000 BP (Fig. 7.1b). A strong warming trend is evident between 10 000 and 8000 BP (Fig. 7.1c) and a cooling trend is shown in the last 4000 years (Fig. 7.1d).

Moisture conditions in the summer and winter rainfall regions are considered separately. Figure 7.2a indicates that throughout the summer rainfall region palaeoclimatic data indicate conditions wetter than the present for much of the last 40 000 years. Clearly, some of this evidence must be considered as 'noise' in the system, and over the last 5000 years at least, it was only slightly wetter than at present. There are two periods with a relatively large number of dates for conditions wetter than at present: 26 000–20 000 BP and 16 000–12 000 BP, with smaller peaks between 7000–6000 BP and 2000–1000 BP. Clear drying trends are evident in the periods 22 000–20 000 BP and 12 000–10 000 BP (Fig. 7.2c). The strongest evidence for a widespread period of conditions drier than the present is for the period 10 000–8000 BP, with a further period of desiccation after 4000 BP (Fig. 7.2b).

In the winter and all-year rainfall region, there are few sites with well dated evidence for moisture changes. The strongest evidence for a period of wetter conditions dates to 14 000–10 000 BP, and again around 4000 BP. It is clear from Fig. 7.3b that conditions were much drier during the period 22 000–16 000 BP.

The bar charts in Figs 7.1 to 7.3 highlight the fact that even primary data in Southern Africa do not provide a clear and unequivocal record for palaeoenvironmental changes, but they do identify the strongest signals. They also show that, although observations are limited, the summer and winter rainfall regions are not consistently out of phase with each other. The relatively crude results from this exercise help to stress the value of palaeoclimatic data based on several different lines of evidence from the same dated sequence. It is from such sequences that we can hope to get the most detailed information on palaeoenvironmental change.

8 Explaining the palaeoenvironmental patterns

Two aspects of palaeoenvironmental research have been explored so far in this book. The first describes the nature, scale, and timing of changes in sedimentological and biological sequences dating to the past 130 000 years in Southern Africa. The second examines the reasons that have been proferred to explain how and why the changes occurred. The results are summarized below, together with remarks on the implications of the observations for biogeography and human population distributions.

8.1 The nature, scale, and timing of change in Late Quaternary palaeoenvironments

All researchers are in agreement that palaeoenvironmental adjustments are directly or indirectly related to the worldwide cycle of glacial and interglacial events of the Late Quaternary. Changing global temperatures were the primary factor in initiating climatic changes that led in turn to adjustments in plant and animal community structure and to disruption of cycles of deposition and erosion. To investigate Southern African climatic and environmental reactions to these fluctuations in temperature we need to be able to interpret accurately the clues that remain in the geological and archaeological record. These interpretations are not always straightforward. By using a combination of inductive and deductive methods we tend to discern patterns, and then try to explain them using modern analogues. To make the explanations believable, it is essential that the interpretations of the field data are critically examined. If this is not done, we may find ourselves explaining events that never happened.

8.1.1 The nature of the terrestrial record

The geological history of Southern Africa in the Quaternary has been

characterized by tectonic stability and the main process involved in landform evolution has been erosion rather than deposition. The subcontinent lacks deep sedimentary basins, and alluvial and colluvial mantles are thin. In addition, seasonal climates and soil pH in general are not favourable for the preservation of plant and animal fossils. These conditions make the compilation of a terrestrial record of palaeoenvironmental changes for Southern Africa a complex task.

Early attempts at reconstructing Late Quaternary palaeoenvironments and their relationship to past climatic conditions were largely in the field of the earth sciences. Geological and geomorphological studies continue to be important, but have been complemented by increasing emphasis on palaeontological, palynological, and stable isotope studies. This has enhanced the resolution of reconstructions and has promoted multidisciplinary studies of the same sequences.

The terrestrial record may be patchy and incomplete, but it nevertheless confirms that the scale and timing of climatic changes in Southern Africa were essentially similar to those in other southern hemisphere countries. Many of the details still need resolution. In particular, methods of interpreting past climatic events from alluvial and colluvial processes need refining; more precise data are needed on the distribution patterns of modern plants and animals, and the degree to which they can tolerate changes in environmental parameters. The interaction between animals, plants, and climate is a dynamic one, and we cannot expect that faunal and floral communities will remain static. Faunal remains, pollen, and charcoal data show that plants and animals do not react to climatic change as wholly integrated communities, but each taxon or alliance has its own range of tolerances and adjusts differently. The nature, scale, and timing of changes in Late Quaternary environments as deduced from field and laboratory studies thus serves the dual purpose of informing on the dynamic changes in biogeographic range and ecosystem history, and of providing a test for models of changes in synoptic climates in the subcontinent.

As we have pointed out, the study of sediments, fossils, and their isotopic composition provide proxy rather than direct evidence for past climates. These diverse data have to be interpreted in terms of climatic change as a common denominator and this is done more easily in a relative sense than in absolute terms. For this reason, the most reliable information comes from the longer, more continuous sequences where changes in environmental conditions as gauged from several different lines of evidence at one particular place can be compared against the standard of the present day. By holding place constant, climatic factors can be more easily evaluated.

8.1.2 The scale and timing of the strongest signals

In assessing 'good' evidence from the not-so-good, the most reliable information surely comes from long sequence sites in which several different lines of palaeoclimatic evidence have been studied from the same dated units. In Southern Africa such sites include caves or rock shelters where deposits have been built up through occupation by Stone Age people, carnivores, or owls. The faunal, floral, and sedimentological data provide cross-checks that help in evaluating the signals from various sources (Beaumont *et al.* 1978; H. J. Deacon 1979; Deacon *et al.* 1984*a*; Beaumont 1986). Equally important are the dated long sequence temperature records now available from the Uitenhage aquifer (Heaton *et al.* 1986*a*) and from speleothems (Talma *et al.* 1974; Vogel 1983), as well as offshore sediments and tree rings which have yet to be widely used. Next in importance are regional climatic events that are well dated at a number of different sites, such as the widespread wet phase between *c.* 40 000 and 22 000 BP in the Kalaharian and adjacent ecozones (Lancaster 1979, 1987; Heine 1981, 1982, 1984; Van Zinderen Bakker 1982*a*; Butzer 1984*a,b*), or a series of pollen sequences from different sites in the Transvaal (Scott 1982*a*, 1984*a*). Most lacustrine (pan) alluvial and colluvial deposits represent one or more discrete episodes, and do not provide the same quality of sequential information as cave sites. Their value is therefore determined by the precision of dating and correlation that is possible in a regional framework, and the reliability with which palaeoenvironmental interpretations can be made. Less useful are isolated but dated observations (for example, Horowitz *et al.* 1978), and least useful are undated or poorly dated isolated observations such as faunal remains from open sites (Klein 1980).

By weighting evidence in this way we can usefully select and summarize the strongest signals that have been highlighted in chapters 4–7. The most obvious contrast is that between full interglacial and full glacial temperatures with a clear difference in all ecozones that could have been as much as 9.5°C in the Transvaal and in the Cape was of the order of 5–5.5°C (Talma *et al.* 1974; Vogel 1983; Heaton *et al.* 1986*a*). Effective precipitation was highest during periods when temperatures were intermediate between full interglacial and full glacial, with a longer period of relatively high rainfall during the lead down to the Last Glacial Maximum between about 40 000 and 25 000 BP when temperatures became gradually cooler, and a shorter period in the period immediately following the Last Glacial Maximum between about 16 000 and 12 000 BP when temperatures rose very rapidly (Cooke 1980, 1984; Heine 1982; Deacon *et al.* 1984*a*; Heine and Geyh 1984; Kent and Gribnitz 1985; Teller and Lancaster 1985; Beaumont 1986). The widespread nature of

these periods of increased rainfall suggests that evaporation rates, circulation patterns, and relative temperatures of land and sea were at their most favourable for precipitation in Southern Africa during the lead down to and out of a glacial maximum. Where biological evidence is available it shows that the Last Glacial Maximum *sensu stricto* was for the most part exceptionally dry, but in the Kalaharian and adjacent ecozones aridity is suggested only by the absence of evidence for higher precipitation. At other times, regional climatic factors played a stronger role and there is some variability in precipitation in different ecozones and under different rainfall regimes during interglacials, and during periods of gradual temperature reduction after interglacials.

The most reliable evidence for conditions during the Last Interglacial comes from Klasies River (Singer and Wymer 1982; Deacon *et al.* 1986). Oxygen isotope measurements show values close to those of the Present Interglacial at the base of the sequence with a significant regression registered during the time the site was occupied by people making Howiesons Poort-type Middle Stone Age artefacts. The scale and timing of this regression suggest that it correlates with oxygen isotope stage 4 and with the sea level regression between Va and IVa recognized by Chappell and Shackleton (1986) on the Huon Peninsula in New Guinea where a regression is dated to between 80 000 and 60 000 BP. An important implication is that this stage of the Middle Stone Age in other similar sequences, such as at Nelson Bay Cave (Klein 1972*a*), Border Cave (Beaumont *et al.* 1978) and Cave of Hearths (Mason 1962), probably also dates in part to the Last Interglacial and to the terrestrial equivalent of oxygen isotope stage 4.

Although biological evidence suggests there was a step-wise temperature reduction with warm and cooler spells during the time period equivalent to stages 5, 4, and 3 of the oxygen isotope sequence (Beaumont *et al.* 1978; Klein 1980, 1986), these events are not precisely dated in Southern Africa, but can be traced in long sequences such as at Klasies River, Border Cave, and Boomplaas Cave where there was an amelioration of temperatures and wetter conditions at *c.* 60 000 BP post-dating the Howiesons Poort or equivalent substage of the Middle Stone Age.

Dune systems in the Kalahari which indicate stronger winds during a drier climate in that ecozone (Lancaster 1981) are undated, but could relate to the early part of the Last Glacial because they became stabilized and vegetated after *c.* 35 000 BP.

The period of highest rainfall in the summer rainfall region (Kalaharian, Karoo-Namaqualian, and Transvaalian ecozones) was between *c.* 40 000 and 25 000 BP when pan and lake levels were substantially higher (Cooke 1975, 1980; Butzer *et al.* 1978*b*; Lancaster

1979; Heine 1981, 1982; Butzer 1984*a,b*). The Boomplaas sequence in the southern Cape also has evidence for higher rainfall than at present at this time, and Scholtz (1986) tentatively suggests from the wood anatomy of charcoal samples that it may have been mostly winter rain. Unfortunately, there is no well dated evidence from the winter rainfall region of the western and south-western Cape for this time period although micromammals at Die Kelders (Avery 1982*a*) indicate moister conditions that could coincide with part of this time zone.

The coldest period of the Last Glacial Maximum is dated to between *c.* 20 000 and 17 000 BP (Talma *et al.* 1974; Vogel 1983; Heaton *et al.* 1986*a*) and where reliable evidence is available, it is accompanied by dry conditions. Exceptions are the Gaap Escarpment and Alexandersfontein (Butzer *et al.* 1978*b*; Butzer 1984*a,b*) and pans in the north-western Cape (Beaumont 1986). Where available, biological evidence from several ecozones shows conditions to have been exceptionally harsh with low species diversity amongst plant and small mammal communities and in some cases plant communities that have no modern analogue (Avery 1982*a*; Scott 1982*a*, 1984*a*; Deacon *et al.* 1984*a*). Xylem analysis of charcoal from Boomplaas, where today rainfall is at all seasons, but with spring and autumn peaks, suggests dry conditions but with a summer rainfall peak (Scholtz 1986). There is again no unequivocal evidence for this time period from the winter rainfall region although sediments from Elands Bay Cave that probably date between about 20 000 and 12 000 BP suggest it was cold and dry (Butzer 1979). The south-western Cape palaeosols described by Butzer (1984*b*) are neither well developed nor well dated. Sea surface temperatures off the western and eastern coasts of southern Africa were 2–5°C cooler than at present and would certainly have lowered the rate of evaporation (Morley and Hays 1979; Prell *et al.* 1979; Martin 1981; Newell *et al.* 1981; Van Zinderen Bakker 1982*a*).

There is some evidence for moister conditions again after the Last Glacial Maximum in the summer rainfall region in the Kalahari, Namib, and north-western Cape (Heine 1982, 1984; Van Zinderen Bakker 1982*a*; Beaumont 1986), and the Boomplaas data show the highest rainfall in the entire sequence between *c.* 16 000 and 11 000 BP (Deacon *et al.* 1984*a*). After *c.* 10 000 BP the relationship between evaporation and precipitation, and between temperatures on land and oceans changed again as ocean temperatures rose (Vincent 1972). In the period between 12 000 and 9000 BP the dating resolution is not quite fine enough in Southern Africa to find widespread support for the short-term temperature fluctuations such as the Bölling, Allerød, and Older and Younger Dryas recognized in Europe, nor for the two-step deglaciation recognized in some deep-sea and polar ice cores (Ruddiman and

Duplessy 1985), although Coetzee (1967) has recognized some of these stages in the pollen sequence at Aliwal North.

Several long sequence sites with good palaeotemperature records (Uitenhage aquifer, Cango, and Wolkberg speleothems) have depositional breaks between the end of the Last Glacial Maximum and the Early Holocene, but there is no evidence from other sources to suggest that temperatures at *c.* 9000 BP were higher than during the mid-Holocene as they apparently were in New Zealand (Salinger 1984) and Australia (Harrison *et al.* 1984). Where details are available, it seems that the highest temperatures in Southern Africa occurred between 7000 and 5000 BP. Holocene temperatures generally fluctuated about 1°C about the present day mean (Martin 1968; Scott 1982*a*, 1984*b*; Vogel 1983; Deacon *et al.* 1984; Heaton *et al.* 1986*a*; Thackeray 1987; Vogel and Talma, personal communication).

Rainfall fluctuations during the Holocene are again of a lower order than has been estimated for changes during the Late Pleistocene. In the Early Holocene, locally moister conditions are reported from the Kalaharian Ecozone (Butzer *et al.* 1978; Heine 1982), from Siphiso rock shelter in Swaziland (Prior and Price-Williams 1985), and from Elands Bay Cave in the winter rainfall region (Klein 1984); but elsewhere a drying trend is indicated, and summer drought is suggested by the wood anatomy of charcoals from Boomplaas (Scholtz 1986) and East Griqualand (Tusenius 1986).

The mid-Holocene record is also patchy with drier conditions indicated from Wonderkrater and Siphiso in the Transvaalian Ecozone (Scott 1982*a*; Prior and Price-Williams 1985), East Griqualand in the Basutolian Ecozone (Tusenius 1986), and from charcoal and micromammals from Boomplaas and Byneskranskop at and after 6400 BP in the Cape Ecozone (Deacon *et al.* 1984; Avery 1982*a*, p. 338). Moister conditions are reported from the Kalaharian Ecozone at Wonderwerk, Gaap Escarpment, and Kathu Pan (Butzer *et al.* 1978*b*; Butzer 1984*a,b*; Beaumont 1986), and from the Cape Ecozone at Groenvlei (Martin 1968).

Somewhat cooler temperatures are recorded during the last 4000 years, but with some warmer temperatures apparent from the Cango speleothem around 2000 BP (Vogel and Talma, personal communication). Transvaalian Ecozone pollen sequences show conditions similar to or drier than those of the present (Scott 1982*a*, 1984*a,b*), although Siphiso and East Griqualand suggest it was relatively moist during this time (Prior and Price-Williams 1985; Tusenius 1986). Cooler and drier conditions are also suggested by a retreat of forest in the southern Cape (Martin 1968; Scholtz 1986), but Boomplaas charcoals of the last 2000 years indicate warm and mesic conditions with a lower incidence of drought than in the Early and mid-Holocene (Deacon *et al.* 1984*a*; Scholtz 1986). In the winter rainfall belt of the western Cape, micro-mammals suggest it was somewhat moister after 3000 BP at Elands Bay Cave and after 2000 BP at Byneskranskop (Avery 1982*a*, 1983). It is apparent that there does not seem to be a consistent pattern during the Holocene between temperature and rainfall fluctuations in the summer and winter rainfall regions. The implication, as Heine (1984, p. 1751) has remarked, is that the variability of rainfall and temperature responses in Southern Africa to global temperature changes shows that regional thermal and hydrologic changes cannot be used as chronostratigraphic reference points. Where detailed biological data are available, it is only within the last 5000–4000 years that floral and faunal communities assumed their present-day character.

8.2 Biogeographic implications

Biological indicators such as plant and animal fossils provide some of the clearest evidence for the scale and timing of past climatic changes in the Late Quaternary, but whereas the effects of climatic forcing on biological communities is readily discernible, it is less easy to explain the processes that may have been involved in these changes. The reason is simply that the role of environmental determinants on the functioning of ecosystems in Southern Africa is generally not well understood.

Recent phenological studies in the fynbos vegetation of the Cape ecozone (Pierce 1984) show that the seasonal rhythms exhibited within a taxon or between taxa in a community are very variable, precluding simple causal explanations in terms of environmental determinants. On the other hand, environmental conditions are important in constraining distributions as has been shown in a study of the phytogeography of the Poaceae (grasses) in Southern Africa where subfamily regions correspond closely to zones defined on modern climatic parameters (Gibbs Russell 1986).

Any change in the length of the growing season, season of precipitation, soil moisture content, or other ecophysiological factors, would affect plant species differentially and would alter the interaction with animals serving as pollinators, with seed dispersal agents, and with herbivores and in turn their predators. In this way there is considerable potential for dynamic changes in biotic communities over time that can be expected to be reflected in the fossil record. Two examples, Boomplaas Cave in the southern Cape and Wonderkrater in the Transvaal, involved different ecosystems, but show parallel and equivalent scales of Late Quaternary changes. They show the strength of the palaeobiological approach in providing data on ecosystem history, or what Poynton (1986) has termed the historical dimension of ecological biogeography.

At Boomplaas the sequence shows changes in biotic communities in the Cango Valley from *c.* 70 000 BP to the present. During this time, charcoal and pollen analyses show that woodland was replaced by treeless karroid vegetation at the Last Glacial Maximum, and in turn by woodland dominated *Olea* in the terminal Pleistocene, by thicket vegetation in the earlier Holocene, and by modern woodland dominated by *Acacia karroo* in the last 5000 years. Similar changes in habitat are seen in the large and small mammals, and there is supporting evidence for the dating and scale of changes in the sediments and stable isotopes (Deacon *et al.* 1983, 1984*a*). The contrast between the biotic communities of the Last Glacial Maximum and those of the later Holocene is dramatic, and is certainly explicable in gross terms as the result of climatic change. The cold and dry conditions at the Last Glacial Maximum favoured communities that were very different in structure and composition from those found in the valley today, and were different also from communities found today at higher altitudes in the surrounding mountains. They were therefore unique combinations of taxa that correlated with unique climatic conditions.

The Holocene record from Boomplaas is important in showing that modern combinations of plant and animal taxa have developed only within the last 5000 years. The implication is that the climate (hot summers, cool winters, and rainfall all-year with peaks in spring and autumn) that plays a major part in determining the present-day range of vegetation, is also the result of factors that have combined only within the last 5000 years. The carbon-13 content of the Cango speleothem (Fig. 6.13) shows a significant change in vegetation in the Late Holocene from plants dominated by C_3 species that grow under cool conditions in the Late Pleistocene and mid-Holocene, *c.* 5500 BP, to plants dominated by summer-growing C_4 species after *c.* 5000 BP. At Boomplaas and Buffelskloof, and at a number of other sites in the southern and western Cape, there is also a shift in the range of larger mammalian species with modern alliances present only within the last 5000 years (Klein 1978*a,b*, 1980, 1984). The larger mammal data, gathered from sites throughout the Cape Ecozone, confirm that this is not a site specific phenomenon.

In the Cango Valley, the replacement in the later Holocene of thicket vegetation by an *Acacia karroo* woodland alliance of C_4 plants is explicable in terms of a relatively subtle climatic change that includes a hot summer and cool moist winter (Cowling 1983*b*, 1984; Pierce 1984; Pierce and Cowling 1984). A change in precipitation and soil-water balance is also indicated in the Late Holocene by the reinitiation of stalagmite formation in the Cango Caves after *c.* 5500 BP. Furthermore, the presence and prominence of the C_4 grass *Themeda triandra* in the Cango Valley together with *Acacia karroo* indicates an alliance of subtropical generalists (Cowling 1983*a*, p. 141). These are taxa that can utilize biseasonal precipitation and, in the case of *T. triandra*, allow it to displace in part the winter-growing C_3 grasses because it can exploit winter and summer growing seasons equally well (Cowling 1983*b*, p. 125). *Acacia karroo* is a species that has been a major invasive in several biomes in the recent past, and although this may have been promoted by the introduction of domestic stock, charcoal studies from Buffelskloof (Deacon *et al.* 1983) show that the initiation precedes domestic stock and is a product of later Holocene environmental conditions. These conditions promote a winter and summer growing season that is different from the Late Pleistocene and Early Holocene climatic regime under which C_3 plants requiring a cool growing season were favoured (Vogel 1978; Cowling 1983*b*; Vogel and Talma, personal communication).

Cowling (1983*a*) has shown that vegetation communities of the south-eastern Cape are sensitive to climatic change, but that different communities would have reacted in different ways to climatic forcing. By correlating available data on phylogeny, speciation and endemism of plant taxa in this region, he has hypothesized that Cape fynbos taxa are limited by edaphic barriers and the flora was therefore never displaced from its present area of distribution during the Pleistocene. Renosterveld, by contrast, would have fluctuated dramatically depending on prevailing climatic conditions, and subtropical thicket and grassland, and Afromontane forest would have been displaced during colder climatic periods and would have become re-established only during warmer and wetter periods. During drier and colder glacial times, karroid taxa would have invaded areas at present occupied by renosterveld and thicket (Cowling 1983*a*, p. 393). This hypothesis again implies that modern climatic conditions correlate with the *Acacia karroo* and *Themeda triandra* alliance, and that hot summers, cool winters, and all-year, but relatively unpredictable rainfall pertained in the southern Cape only within the last 5000 years. Dynamic reconstructions of vegetation history such as this one will be essential if we are to use palaeontological and palaeobotanical studies in more than a descriptive way to evaluate the complex interrelationships of palaeoclimate and biogeographic factors.

The second example is the Wonderkrater spring deposit in the Transvaalian Ecozone where the pollen sequence has been studied by Scott (1982*a*). As in the Cango Valley, woodland of a different character from that of the present grew around the site prior to the Last Glacial Maximum and was replaced by ericaceous heathland taxa between *c.* 25 000 and 11 000 BP. In the period *c.* 11 000–4000 BP savannah vegetation was dominant and it was only from 4000 BP to the present that the

modern bushveld taxa were present. Scott (1982*a*) has interpreted this sequence of changes by relating the palaeocommunities to analogues along modern altitudinal and moisture gradients, and on this basis to infer past climatic conditions (Fig. 5.8), but he has also noted that the palaeocommunities are not exactly the same as the modern analogues. With the knowledge that there are no exact modern analogues for past vegetation communities, it will be advantageous to undertake phytochorological studies of the Wonderkrater and surrounding flora before drawing detailed climatic inferences, but it is nevertheless important to note that the modern bushveld alliance is only about 4000 years old and, by implication, so is the present-day rainfall and temperature regime.

Present-day ecosystems are the product of biological processes of evolution, speciation, dispersal, and extinction. The association of plant and animal species in communities is not fixed, but is the product of selection operating over time. The research summarized in this book has shown that climatic changes in the Late Quaternary have been a pervasive factor in determining the composition and nature of modern natural communities, and that modern distributions were achieved only in the last 5000–4000 years. However, during the last 2000 years and particularly since the advent of European colonists, the natural communities have been considerably modified by land use patterns. These agricultural and settlement patterns have lowered the diversity of ecosystems, and have reduced their resilience simulating conditions more similar to the Last Glacial than to the Late Holocene. This loss of diversity through extinction or range restriction of taxa is serious because it implies that as Southern Africa moves into the next glacial, communities will be impoverished to an even greater extent than before. The goals of conservation and management must therefore be directed at maintaining diversity in general, at restoring the natural range of taxa, and at reducing the risk of the loss of endangered species. Whereas processual studies on the functioning of modern ecosystems have received welcome attention in the last decade in Southern Africa, there is an equal need to invest research resources in historical studies.

8.3 Implications for human population distributions

As the data summarized in this book have shown, climatic changes of the Late Quaternary induced significant changes in plant and animal distributions, and in turn these affected human habitats, particularly in the availability of food resources. The distribution and density of human populations was constrained by the ability of people to cope with changes in the natural productivity of their environments because prior to the last 2000 years, agriculture was not practised in Southern Africa.

Africa as a whole is a continent of seasonal rainfall, except for the limited area of the equatorial forest biome, the Guinean Ecozone, and, like Australia, is generally arid and prone to episodic droughts. Persistent droughts lasting many years have been recorded historically, and in regions such as the Sahel and Ethiopia are the cause of famine and stress, as well as of high local extinction rates. Demographic studies among the !Kung San in Botswana (Howell 1979) indicate that fertility is linked to environmental conditions through fat levels and lower birth rates are a response to a reduction in food resources. In this way there is a homoeostatic control on population in accord with natural productivity.

In a study of the archaeological evidence for fluctuations in human population levels during the Late Quaternary, H. J. Deacon (in preparation) suggests that it was Late Pleistocene aridity rather than reduced temperatures that was primarily responsible for lower population levels in Southern Africa. Archaeological visibility (measured by the number of Stone Age occupation sites) was relatively high in Middle Stone Age times during the Last Interglacial, and was high again during the Later Stone Age in the Present Interglacial. In most long sequence sites in all ecozones, discontinuities that indicate significant time breaks separate Middle and Later Stone Age occupation horizons, and Last Glacial occurrences are rare.

Figure 8.1 (from H. J. Deacon, personal communication) summarizes the number of radiocarbon dates that have been published for archaeological deposits in various regions of Africa. They date within the past 40 000 years when the method is applicable and are expressed on a relative scale (RND) for the continent so that they can be compared at the same scale even though the actual number of dates varies from region to region. The assumption is that the radiocarbon dates represent the frequency with which archaeologists have intersected human occupation horizons during excavations and in turn this represents a crude indication of the density of people in the landscape at any particular point in time. In all regions there are relatively few occurrences older than 10 000 years, but the Atlas Region, the Sahara, North-east Africa and, to a lesser extent, East Africa show a significant increase in dated horizons from about 10 000 BP that correlates with a period of markedly higher precipitation and a stronger monsoon system in the Early Holocene (Nicholson and Flohn 1980) evident from higher lake levels (Street and Grove 1976). West Africa, the Miombo Region of Central Africa, and Southern Africa, on the other hand, show a significant increase only in the last 5000 years suggesting they were affected by different climatic factors that brought drier conditions during the Early and mid-Holocene, and higher moisture levels only within the last 5000 years (Martin 1968; Scott 1982*a*, 1986; Butzer 1984*a*,*b*; Deacon *et al.* 1984*a*;

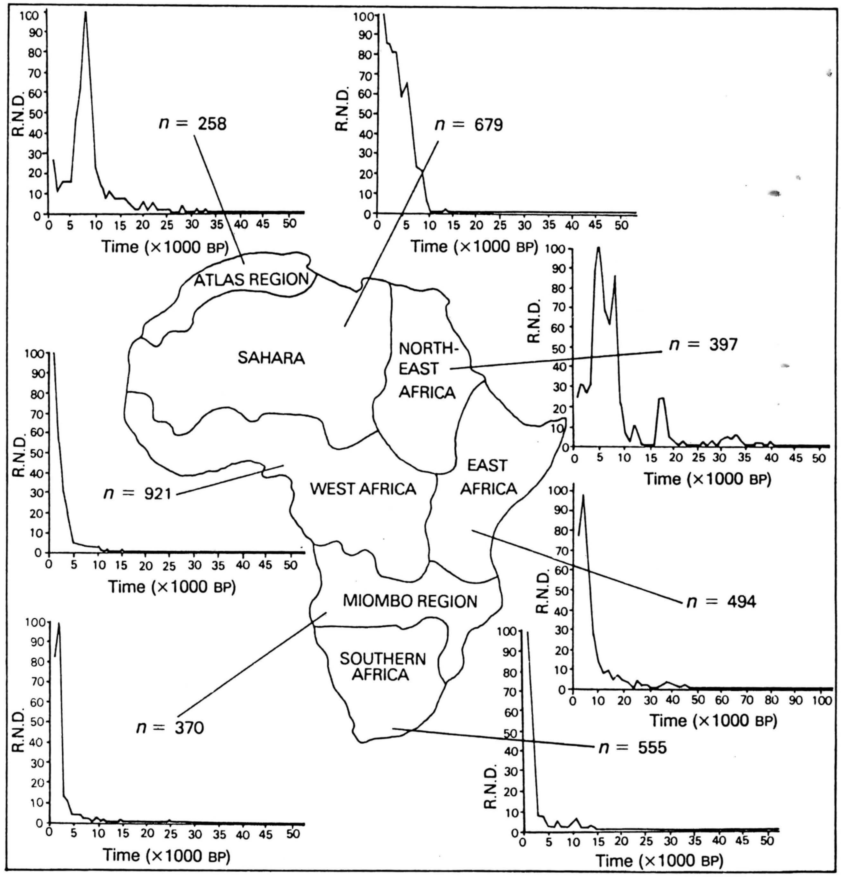

Fig. 8.1 Relative number of radiocarbon dates in various regions of Africa showing that North and East Africa are out of phase with the rest of the continent (from H. J. Deacon and J. F. Thackeray, personal communication).

Prior and Price-Williams 1985; Scholtz 1986; Tusenius 1986). As Nicholson and Flohn (1980) and others have noted, the Southern African precipitation regime is out-of-phase with that to the north and east.

Throughout Africa, the increase in dated archaeological occurrences precedes the appearance of food production by several thousand years suggesting that the spread of food production (which originated earlier in North and Eastern Africa than in West and Southern Africa) was a consequence rather than a cause of Holocene population increases. On the other hand, food production has allowed populations to climb to density levels far beyond those that could have been maintained by hunter-gatherer subsistence systems alone (H. J. Deacon, personal communication).

8.4 Implications for palaeoclimatic models

Climatic models proposed to explain regional changes and variations in rainfall have been refined over the past 25 years (see chapter 2.2) and the most recent ones by Nicholson and Flohn (1980), Harrison *et al.* (1984), Tyson (1986), and Cockcroft *et al.* (1987) have made more sophisticated use of modern climatic analogues. The former two explore the implications of movements of the ITCZ south of its present position because of a much larger extent of continental ice in the northern hemisphere at the Last Glacial Maximum. For Southern Africa their model suggests more vigorous westerlies situated further north than today, but a reduction in precipitation, especially over land during the Last Glacial Maximum (Harrison *et al.* 1984, p. 29). The latter two models, on the other hand, suggest that Last Glacial Maximum climates were dominated by an enhanced Southern Oscillation that would have brought dry conditions to the summer rainfall region and wetter conditions to the winter rainfall region. Winter rains would have extended to 25° south and well inland of present limits. They assume that during periods of cold temperatures the Southern Oscillation would have been in a low phase (i.e. summer drought in the summer rainfall region) and that during periods of warmer temperatures the Southern Oscillation would have been in a high phase bringing increased rainfall to the summer rainfall region. During a low phase, westerly winds and cyclones would have been drawn northwards increasing winter rainfall, and during a high phase the opposite would occur bringing winter drought. As explained in section 2.2.3, the test implications are that Last Glacial Maximum deposits should record higher rainfall north and east of the winter rainfall region. Dry periods in the summer rainfall region should coincide with wet periods in the winter rainfall region and vice versa, should this explanation be correct.

There is a difference between the two models, although both predict an equatorwards movement of the westerlies and winter rains. Harrison *et al.* (1984) suggest Last Glacial Maximum rainfall was probably similar to or lower than at present in the modern winter rainfall area, with Nicholson and Flohn (1980) suggesting increased winter rain to the north of this area (Fig. 2.7). Tyson (1986) and Cockcroft *et al.* (1987), on the other hand, predict higher rainfall from the westerlies both in the present day winter rainfall region, and to the north and east. An important corollary of their argument is the negative correlation between

the summer and winter rainfall regions. Neither of these predictions is convincingly met in the data summarized here, mostly because there are so few well dated observations for past climates in the winter rainfall region. However, what little evidence there is (see chapter 6.6.4), supports Harrison *et al.* (1984) in suggesting that conditions in the surrounds of Elands Bay Cave in the western Cape at the Last Glacial Maximum, where deposits are dated within the broad limits of *c.* 20 000 to 11 000 BP, were as cold and dry as they were elsewhere in both summer and all-year rainfall regions (Butzer 1979; Parkington 1986). Although wetter conditions are reported in the north-western Cape and in the Kalaharian Ecozone during the Last Glacial Maximum (Butzer 1984*a,b*; Beaumont 1986), there is no way of testing yet whether this was the result of summer or winter rainfall. The evidence against the latter is the lack of change to C_3 grasses that would be expected at Apollo 11 at 28° south had winter rainfall penetrated to 25° south during the Last Glacial Maximum (Vogel 1983). The only technique that has been developed so far to indicate season of rainfall is xylem analysis used on charcoal samples from Boomplaas and this tentatively suggests dry conditions, but summer rainfall during the Last Glacial Maximum in a region that today receives rain all year round (Scholtz 1986). The fact that the southern Cape mountains have no block streams and that no true eboulis sec is found in caves in the region also suggests that the season of maximum cold did not coincide with a rainfall peak and argues against higher winter rainfall during the Last Glacial Maximum.

We have to conclude that the most reliable palaeoenvironmental evidence available in Southern Africa does not support a model in which the summer and winter rainfall regions were out of phase in the last glacial cycle. On the other hand, this is not to say that the model may not be proved correct when there are more data from the winter rainfall region. At present, we simply do not have enough convincing evidence to support or reject the model outright. As Heine (1984) has shown, combined data from Southern Africa indicate that higher rainfall occurred in the 15 000 years before and in the 4000–5000 years after the Last Glacial Maximum when temperatures were about 2–4°C cooler than at present. During both colder and warmer periods than this, the effect was more variable in different regions of Southern Africa and no clear pattern has emerged yet. More importantly, the historical biogeographic implications of the analyses of biological data indicate that modern communities of plant and animal taxa developed only within the last 5000 years when human population densities in Southern Africa suggest productivity was higher than during the Last Glacial and Early Holocene. This observation in turn implies that modern climatic patterns, including the degree of seasonality that existed between summer and winter temperature, and rainfall regimes, and by the same token the boundaries between winter, summer, and all-year rainfall zones, also date within the last 5000 years. If this is indeed the case, then modern climate analogues may be inappropriate for modelling palaeoclimates for periods older than 5000 years ago in Southern Africa. The challenge for future research will be to test these implications by enlarging the palaeoenvironmental data base and by extending modern studies to provide viable analogues that have greater power to explain past patterns of plant and animal distributions, and the climatic conditions that affect sedimentary sequences. Biogeographic studies, in which environmental changes from a number of different lines of evidence can be integrated, have this potential.

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