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Climate change in arid environments: revisiting the past to understand the future

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Abstract: Arid regions are expected to undergo significant changes under a scenario of climate warming, but there is considerable variability and uncertainty in these estimates between different scenarios. The complexities of precipitation changes, vegetation–climate feedbacks and direct physiological effects of CO_2 on vegetation present particular challenges for climate change modelling of arid regions. Great uncertainties exist in the prediction of arid ecosystem responses to elevated CO_2 and global warming.

Palaeodata provide important information about the past frequency, intensity and subregional patterns of change in the world's deserts that cannot always be captured by the climatic models. However, it is important to bear in mind that the global mechanisms of Quaternary climatic variability were different from present-day trends, and any direct analogies between the past and present should be treated with great caution. Although palaeodata provide valuable information about possible past changes in the vegetation–climate system, it is unlikely that the history of the world's deserts is a key for their future.

Key words: arid lands, climate change.

I Introduction

Deserts and semi-deserts are the most extensive of the world's land biome types, occupying more than 30% of the Earth's surface (United Nations Environment Program (UNEP), 1997). They are often predicted to be among the most responsive ecosystems to global climatic change (Mellilo *et al.*, 1993; Bazzaz *et al.*, 1996; Huxman and Smith, 2001; Whitford, 2002). However, there are still major uncertainties regarding the

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potential effects of increasing concentrations of either CO_2 or future climate change in arid ecosystems. Interpretation of some palaeoclimatic data as the analogues to future climates suggest that global climate change may cause many arid regions to experience higher rainfall and therefore to become more productive ecosystems. It is difficult to say, however, whether there are any valid analogues between the climate changes of the past and those of the future induced by greenhouse gases.

The results of general circulation models (GCMs) in relation to arid environments under a future 'greenhouse effect' climate are complex and contradictory (Hulme *et al.*, 1999; Hulme, 2001; Intergovernmental Panel on Climate Change (IPCC), 2001). Despite the great uncertainties about the responsiveness of arid ecosystems to the ongoing climatic changes, scenarios predicting increases in precipitation in present-day deserts are sometimes interpreted as indicators of a likely increase in productivity of arid zones as a result of the CO_2 increases (Payten, 2000; Batchelor, 2002). Such predictions, however controversial they may be, are often used by policy makers and groups opposed to the greenhouse gas emission regulations, suggesting that global warming will likely enhance the agricultural potential of the arid zones, turning them into a sort of 'greenhouse paradise'.

Evaluation of responses of arid environments to global climate change requires further data collection, experimental work, modelling and interdisciplinary exchange to improve our understanding of climate–ecosystem interactions at various spatial and temporal scales.

The goals of the present review are two-fold:

1. to summarize and discuss the progress in and challenges of climate change modelling and forecast in arid and semi-arid ecosystems, and
2. to discuss the results of palaeoecological reconstructions and model scenarios for four major arid regions (the Saharo-Arabian, Australian, Turanian and the US deserts and semi-deserts). The choice of these regions was dictated by my goal to cover the maximum variety of arid climates, including both tropical and temperate deserts and semi-deserts.

The next section of this paper addresses the Late Pleistocene and Holocene changes in the Saharo-Arabian deserts, Central and Western Australia, Turanian deserts of Central Asia and deserts and semi-deserts of the western USA. The regional palaeodata were used to attempt to explain the effects of global climate change in the last glacial maximum (LGM) and the Holocene optimum on the ecosystem changes in arid zones. Section III discusses some major sources of uncertainties and contradictions affecting climate change predictions and modelling in arid zones.

Finally, Section IV discusses some challenges of climate change modelling in arid regions and some limitations of palaeoreconstructions as a tool for understanding current global climate change. It also discusses some contradictions between the existing biogeography model scenarios and the recent experimental data on the direct responses on desert and semi-desert vegetation to the direct physiological effects of CO_2.

II Lessons from the past

Drastic changes in desert climates in the recent geological past are well documented for many arid regions of the world (Varuschenko *et al.*, 1987; Street-Perrott *et al.*, 1990;

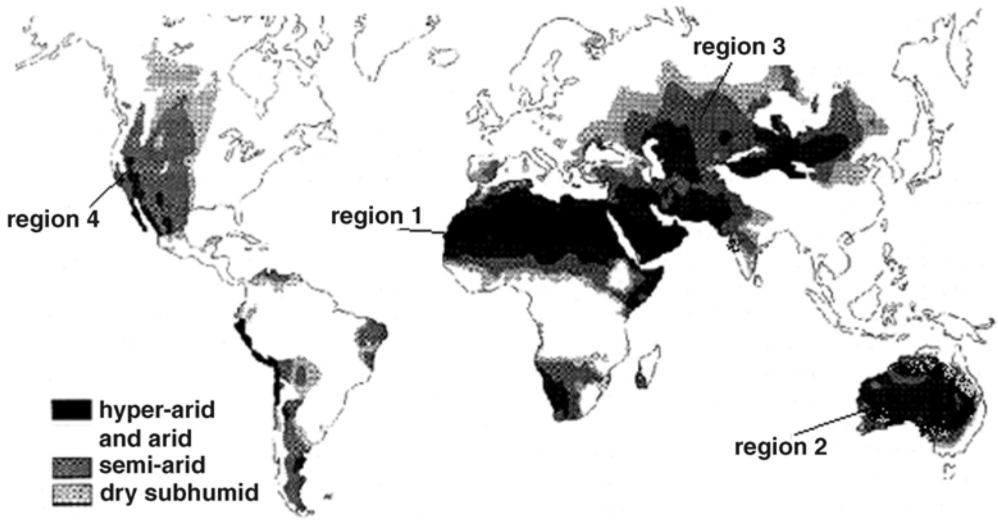

Figure 1 Global distribution of arid zones and location of the four regions discussed in this paper

Nanson *et al.*, 1992; Thompson *et al.*, 1993; Petit-Maire and Guo, 1998). The climate of the present-day deserts and semi-deserts is known to have changed at various temporal and spatial scales. Pleistocene climate variations had a marked effect on many presently arid zones. But can the palaeoreconstructions provide a key to understanding the present trends in arid climates?

1 The Saharo-Arabian deserts

There is abundant information about palaeoenvironmental changes in the Saharo-Arabian desert region during the Late Pleistocene and Holocene, including geomorphic, biostratigraphic and archaeological data. Progress in the understanding of climate and vegetation changes in this region has resulted in a number of regional meso-scale reconstructions and palaeoclimatic/palaeoecosystem datasets during recent years (Sanlaville, 1992; Lioubimtseva, 1995, 1999; Petit-Maire *et al.*, 1995; Hoelzmann *et al.*, 1998; Jolly *et al.*, 1998; Guo *et al.*, 2000).

During the LGM, around 21 000 years BP, the 100-mm isohyet shifted south to 13–14°N in Africa and on the Arabian peninsula (Sanlaville, 1992; Petit-Maire *et al.*, 1995; Petit-Maire and Guo, 1996; Hoelzmann *et al.*, 1998). For comparison, its present latitude is 17–18°N. In contrast, the early to mid-Holocene world around 9000–4000 years BP had much less desert in certain areas. For much of this time the Sahara desert had virtually disappeared and was covered by a mosaic of scrub, grasslands and woodlands populated by giraffes, elephants and other animals that now only survive far to the south. However, this generally moist period seems to have been punctuated by more arid phases (some of which were as dry as at present) often lasting hundreds of years. During the Sahara climatic optimum of 8500–6500 years BP there was an almost 50-fold increased precipitation (by 200–300 mm) compared with the present (Petit-Maire *et al.*, 1995). The Saharo–Sahelian boundary shifted at those times to 23–23°N, that is by 500 km to the north compared with its present-day position, and by 1000 km compared with the LGM situation (Petit-Maire *et al.*, 1995). Considerable increases in precipitation occurred on the Arabian Peninsula, resulting in the total disappearance of arid landscapes and in the spread of steppe and savanna on the Arabian peninsular (Sanlaville, 1992).

More recent but less dramatic fluctuations of precipitation in the Sahara and Sahel, documented and discussed by Nicholson (1994, 2001), suggest that a variety of global and regional factors could cause significant climatic instability in this region during the second half of the Holocene.

Despite the general agreement that climatic variations in the Saharo-Arabian region were triggered by the changes in the orbital parameters of the Earth, it is still unclear what was the role of ecosystem distribution and greenhouse forcing in climatic change in this region. The recent work by Zeng and Neelin (1999), Claussen *et al.* (1999) and Kubatzki *et al.* (2000) suggest that palaeoclimatic changes in the Saharo-Arabian region were triggered by changes in insolation and amplified by a positive, nonlinear biogeophysical feedback between vegetation, atmospheric motion and precipitation.

Our previous studies based on palaeoenvironmental reconstructions of this region (Lioubimtseva *et al.*, 1996, 1998; Lioubimtseva, 1999) showed that during the LGM, reduction of the savanna zone along the southern Saharan margin amounted to approximately 3.3×10^6 km^2 compared with the present, while the area of deserts and semi-deserts increased respectively to 2.2×10^6 km^2 and 5.3×10^6 km^2 northward and southward of their present-day position.

By contrast, climatic changes during the Holocene optimum led to a considerable increase in humidity and spread of subhumid ecosystems. In the Saharo-Arabian region, the area of desert did not exceed 1.7×10 km^2 and its reduction, compared to the present, was approximately 4.2×10^6 km^2 (Lioubimtseva *et al.*, 1998). These results are in relatively good agreement with estimations by Hoelzmann *et al.* (1998) and Jolly *et al.* (1998).

Such changes in ecosystem distribution mean considerable variations in biomass and land-surface parameters (Table 1) that may be responsible for significant atmosphere–vegetation feedbacks.

The method of estimation of carbon storage in vegetation and soils, shown in Table 1, is explained in Lioubimtseva *et al.* (1996, 1998).

The change in biome distribution pattern and latitudinal shifts of vegetation belts resulted in significant variations of carbon storage in this present-day arid region. The computation of the carbon content of the main vegetation types from a reconstruction of the LGM and Holocene has shown that deserts tend to be a sink or source of carbon in response to climatic change. If we consider these palaeoenvironmental changes in terms of carbon storage variations, they would mean that the increase in organic carbon in this region from the LGM to the Holocene Climatic Optimum amounted to about 120 Gt (almost three times). The maximum carbon storage in this region was reached under relatively humid climatic conditions around 9000–6000 years BP. Since the mid-Holocene, increasing aridity caused the release of about 110–146 Gt of organic carbon from the Saharo-Arabian arid region. Careful studies are still needed to assess the biomass decrease from the Holocene optimum to the present with more accuracy, but it is assumed here that it could have decreased approximately 50% or more (Lioubimtseva, 1999).

Table 1 The palaeolandscapes of the Saharo-Arabian region

Ecosystem	Aridity index (P/PET)	Albedo range (%)	Leaf Area Index (range)	Vegetation carbon (kg m^{-2})	Soil organic carbon (kg m^{-2})	Last Glacial Maximum		Holocene Optimum	
						Area ($km^2 \times 10^6$)	Totals of carbon storage (Gt)	Area ($km^2 \times 10^6$)	Totals of carbon storage (Gt)
Extra-arid tropical desert	<0.05	40–90	0–0.25	0.01	0.1–0.5	0.74	0.13–0.38	–	–
Arid tropical desert	0.05–0.2	30–70	0.01–1	0.05–0.2	0.3–1.0	9.97	6.0–11.96	2.1	1.26–2.52
Semi-desert or steppe	0.2–0.5	10–80	0.25–4.0	0.6–1.5	0.9–3.0	1.07	6.52–7.49	1.1	6.16–8.8
Savanna	0.2–1.0	17–50	0.6–4.0	2.5–3.5	2.0–5.5	1.14	14.25–17.67	6.9–8.0	126.35–146.95
Semi-arid Mediterranean woodland	0.5–1.0	14–20	1.0–4.0	4.0–6.0	5.5–7.0	1.05	12.6–16.8	0.76	9.12–12.16

Source: compiled from Lioubimtseva (1995); Lioubimtseva *et al.* (1998); Hoelzmann *et al.* (1998).

The increase of carbon flux to the atmosphere could result both from the abrupt increase of aridity during the second half of the Holocene and from increasing anthropogenic disturbance. On the other hand, both factors had a significant impact on the land surface parameters and biota carbon pools that probably amplified the aridity trend of the second half of the Holocene. Such results of order-of-magnitude biomass changes in arid zones are important in understanding the behaviour of the carbon cycle since the last interglacial. The general increase in the extent of deserts at the expense of arborescent and tall-grass vegetation during the LGM in tropical areas, may have been driven not only by cooling or by drier conditions but also partly by the lowering of atmospheric CO_2.

2 Australian deserts

Much of Australia is currently arid or semi-arid, but with almost no areas of 'extreme desert', such as occur in the Saharo-Arabian region. However, during the LGM most of Australia did indeed have an extreme desert climate, with extensive areas of mobile dunes extending into what are now wooded areas (Kershaw *et al.*, 1991), and dry playa lakes (Nanson *et al.*, 1992, 1998; Cupper *et al.*, 2000). This last major desert phase seems to have begun by around 25 000 years BP, and to have ended sometime before 12 000 years BP. Around the moister periphery of Australia, temperatures seem to have been around 3–5°C lower than at present.

Analyses of pollen records from playas in southwestern New South Wales by Cupper *et al.* (2000) indicates aridity at 7000–6000 years BP, although lakes were at their highest during the mid-Holocene. The presence of *Allocasuarina luehmannii*, a woodland taxon now only found in higher rainfall zones, shows that the high groundwater levels correspond to an increase in local precipitation. Woodland contracted after 4000 years BP with *Casuarina* disappearing from the flora. The Holocene was most arid from 3000–1000 years BP, when some playas dried and wind eroded their exposed floors. Aeolian sand was deposited in the lakes that persisted during this phase. Their pollen records show a further decline of woodland taxa, particularly of drought-sensitive native pine.

Thermoluminescence ages from a longitudinal dune field in northern tropical Australia suggest that complete dune activation occurred here either continuously or sporadically between approximately 8200 years BP and 5900 years BP (Nott *et al.*, 1999). This period, in Australia, is normally ascribed increasingly warm and wet conditions towards the Holocene Climatic Optimum. However, elsewhere, this time (~8000 years BP) coincides with a brief period of global climate change recognized in $\delta^{18}O$ records from Antarctica, methane records of the Greenland Summit ice cores, changes to deep-sea benthic foraminferal composition and atmospheric $\delta^{14}C$ variations. In tropical Africa two distinct phases of aridity have been dated at approximately 8000 and 6000 years BP. The coincidence of aeolian reactivation episodes in this north Australian dune field with brief phases of early Holocene climate change elsewhere suggests possible global climatic teleconnections at this time (Nott *et al.*, 1999).

Lake data from the adjacent areas of southeastern Australia suggest that from 7600 to 5500 years BP, there was a period of more modest aeolian activity characterized by 'reduced wind strength or fewer storms' (Stanely and De Deckker, 2002) that corresponds with the time when lakes in southeastern Australia experienced some of their highest levels. Since that time, the authors report that 'the last 5500 years of the record saw a progressively increasing intensity of aeolian activity', saying they 'interpret this change of climatic conditions for the latter part of the Holocene as a "downgrading" of climatic stability with the predominance of wet and dry periods, such as we know them today and that are strongly influenced by the El Niño Southern Oscillation (ENSO)', which they say was 'less effective' prior to 5500 years BP (Stanely and De Deckker, 2002).

Most proxy data suggest that during the mid-Holocene the Australian continent was generally warmer during winter but that the summer was cooler than today (Kershaw *et al.*, 2000; Moss and Kershaw, 2000). Palaeomodelling scenarios of the Holocene by the University of Melbourne GCM (MUGCM) also confirm that relative to the present day Australian surface air temperatures were lower during the Austral summer and autumn and higher during the winter and spring months (Simmonds, 2003). As a result, there was probably a reduction in the strength of the summer monsoon and its associated precipitation. However, the reduction in rainfall directly associated with the monsoon in the north was more than offset by enhanced summer precipitation associated with a greater number of tropical cyclones in the simulation. In general, Australia experiences less precipitation during each season, with the exception of spring. A consequence of the precipitation response, integrated over a year, is that the modelled Australian surface moisture content is drier than today. In contrast, northeastern Queensland, a location of many palaeo studies, is wetter, because of the enhanced precipitation of the region averaged over an annual cycle.

3 The Turanian deserts of central Asia

While relatively limited biostratigraphic and geomorphological data are available on the Late Pleistocene and Holocene of the Caspian coast (Kes *et al.*, 1993; Varushchenko *et al.*, 1987; Velichko *et al.*, 1987), Aral Sea (Sevastianov *et al.*, 1991) and mountain ranges of Tian Shan and Pamiro-Alaï (Sevastianov *et al.*, 1991), there are virtually no published palaeogeographic data from the desert and semi-desert areas of Uzbekistan and Turkmenistan. There are only a few sites in the steppe and forest-steppe zones of northern and central Kazakhstan where palynological analysis of palaeolake deposits (Aubekerov *et al.*, 1989; Tarasov, 1992) has provided information about the relatively humid early Holocene palaeoclimate and ecosystems in this region. However, the assumption, based on this dataset (the only one available in the steppe zone of Kazakhstan of similar climatic dynamics in the Holocene throughout much bigger arid areas southward by several authors (Peyron *et al.*, 1998; Tarasov *et al.*, 1998) may be misleading. Even fewer data are available for this region for the LGM. For example, Tarasov *et al.* (2000) assume that a desert climate and biome dominated the entire former USSR Central Asia based on pollen data from the only available site near the high-altitude Lake Chatirkol (3536 m above sea level) in the Tian-Shan Mountains. Hence, the reconstruction by these authors should be regarded as very tentative.

Available biostratigraphic, geomorphologic and archaeological data suggest that a cold arid phase occurred in Central Asia during the LGM. Dating of lateral moraine ridges in the Tian Shan mountains recorded the maximum glacier advances between

19 000 and 20 000 years BP and the most recent advance during the Younger Dryas around 11 500 years BP (Hetzel *et al.*, 2002). In central Kazakhstan, annual precipitation decreased during the LGM by 100–150 mm, mean July temperature was 2°C lower than at present, while the mean January temperature dropped by 12°C compared with the present (Aubekerov *et al.*, 1989). According to pollen and archaeological data in western Turkmenistan, the mean annual temperature decreased by 4.5°C (Kes *et al.*, 1993).

The mid-Holocene was associated with an increase in precipitation in the Kyzyl-Kum desert, where the Holocene climatic optimum is known as the Lavliakan humid phase and has been dated by radiocarbon in different archaeological sites from 8000 to 4000 years BP, with a maximum around 6000 years BP (Mamedov, 1990). Such climatic conditions favoured the development of *Artemisia* and *Gramineae* steppes on the currently desert Usturt plateau (Varushchenko *et al.*, 1987). Marine fossils, relict shore terraces, archaeological sites and historical records point to repeated major recessions and advances of the Aral Sea during the past 10 000 years. Until the present century, fluctuations in its surface level were at least 20 m and possibly more than 40 m. Significant cyclical variations of sea level during this period resulted from major changes in river discharge into it caused by climatic alteration and natural diversions of the Amu-Darya River away from the Aral sea (Kes *et al.*, 1993). According to available pollen data in northern Kazakhstan, the dry cold steppes of the Younger Dryas were replaced in the Holocene by mesophytic forest-steppe vegetation with a maximum increase of arborescent species (Aubekerov *et al.*, 1989; Tarasov, 1992). These changes in precipitation and vegetation cover resulted in a decrease of erosion and favoured soil accumulation processes. However, the existing data suggest that these environmental changes most likely had relatively small amplitudes (100–150 mm or less) compared with some other desert regions in the world (i.e., the Saharo-Arabian deserts).

Based on analyses of historical documents and archaeological data in western Turkmenistan and Kazakhstan, Varushchenko *et al.* (1987) came to the conclusion that precipitation was slightly higher than today between the Caspian and Aral Sea between the ninth and fourteenth century AD, followed by the present-day conditions. It is unclear if the increase of aridity in the fourteenth century was caused primarily by internal factors (albedo and water-balance changes resulting from human impact) or by external global-scale processes. The dry phase in western Turkmenistan identified by Varushchenko *et al.* (1987) apparently coincides with the Little Ice Age, well documented in many other parts of the world.

It is important to bear in mind that the mechanism of atmospheric circulation over the Turanian deserts is different in many ways from those controlling precipitation in the tropical deserts. The atmospheric processes over European Russia and Siberia appear to be the main factors controlling precipitation on the Turanian plains. Analyses of available palaeoenvironmental data and the AGCM scenarios suggest that the early Holocene increase of humidity in Kazakhstan and the northern deserts of Central Asia was caused primarily by the southward shift of the westerly cyclonic circulation, pushed at that time by a strong high-pressure area over the periglacial zones of southern Russia, rather than by the Asian monsoon. Another important control on precipitation change in the Turanian deserts is the level of the Caspian Sea and the rate of evaporation from its surface (Varushchenko *et al.*, 1987). The impact of the Caspian Sea on precipitation in this region creates a very strong connection between the climate of Central Asia and that of European Russia, because the level of water in the Caspian Sea entirely depends on climatic conditions and run-off in the basins of the Volga and Ural Rivers. On the other hand, the level of the Aral Sea (another important factor controlling the temperature and precipitation pattern in this region) entirely depends on the run-off of two major Central Asian rivers: Syr-Darya and Amu-Darya starting in the Pamir and Tian-Shan mountains, and therefore on the rhythm of the mountain glaciations. Finally, the climate of the southern subtropical subregion (the Kara-Kum desert and piedmonts of the Kopetdag Mountains and Pamiro-Alai) is also strongly affected by the Asian monsoon. Currently, westerly cyclones of the temperate zone change their trajectories in summer over the Aral Sea from a west–east to a north–south direction and approach the zone affected by the Indian monsoon over the Zagros mountains. It is very likely that humid conditions in the southwest that occurred between 9000 and 8000 years ago and again around 6000–5000 years BP could have been caused both by the increase of activity of the Indian monsoon that at that time could freely reach areas northward from the low Kopetdag mountains, and intensification of the temperate cyclones.

4 North American deserts and semi-deserts

The history of climate change in the deserts and semi-deserts of the US west and southwest has been studied extensively using a diverse range of indicators of past vegetation and past climates including sedimentology, pollen and packrat (*Neotoma* spp.) middens (Thompson *et al.*, 1993; Bartlein *et al.*, 1998; Thompson and Anderson, 2000). Because of the very patchy mosaic in climate and vegetation zones of the American west, caused by high elevation and extremely complex orographic conditions, coarse-resolution climate models can hardly capture the spatial variability of climate trends in this region with sufficient confidence. Palaeoreconstructions provide very complex spatial patterns of vegetation change in this region that reflect the local effects of topography.

During the last Ice Age, the southwestern USA was much wetter and cooler than it is today. The massive Laurentide Ice Sheet that covered most of Canada and much of the northeastern USA apparently had a great effect on atmospheric circulation and pushed the westerlies far to the south relative to today. Large lakes were present across the Great Basin and in parts of the southwest (Thompson *et al.*, 1993). Pollen and packrat midden data indicate that the LGM vegetation was very open. In the Great Basin and surrounding areas, the LGM open conifer woodlands were characterized by pines, juniper and steppe plants (Thompson and Anderson, 2000) and many of the modern desert species apparently could not live in this region and were presumably displaced southward to warmer climates.

Wetter-than-modern conditions continued into the early Holocene, although the cause was different than during the LGM. During the period of maximum summertime solar radiation the summer monsoon of the southwest was apparently much stronger than it is today. This enhanced circulation apparently brought sufficient rainfall and cloudiness into the southwest to allow cool-adapted species to survive in the modern deserts. However, there is also evidence that timberlines were higher than today at (or near) this time, suggesting warmer-than-present summers at high elevations (Thompson and Anderson, 2000). Climatic conditions in the southwestern deserts became warmer and drier by the middle Holocene, and many of the present-day desert species have become established within their modern ranges over the past 6000 years (Thompson *et al.*, 1993; Thompson and Anderson, 2000).

The mid-Holocene data from this region suggest that relatively little climatic change occurred here over the last 6000 years. While most GCM scenarios predict today a strong increase in precipitation and a decrease of desert areas in the western USA under the current warming trend (IPCC, 2001; US Global Change Research Program, 2001), palaeoevidence suggests a rather complex distribution of climate conditions in the mid-Holocene, with some sites wetter than today (i.e., northeastern Arizona) and some sites considerably drier than at present (Great Basin and Sierra Nevada) (Thompson and Anderson, 2000). Studies of aeolian stratigraphy in the western USA indicate several phases of pronounced aridity and aeolian activity from 7545 to 7035 years BP and from 5940 to 4540 years BP (Gaylord, 1990 in Goudie, 1994). Several studies point to the contrast between drier conditions in the southwestern deserts and wetter conditions in the southern Rocky Mountains in the mid-Holocene (Thompson *et al.*, 1993; Bartlein *et al.*, 1998). These appear to reflect an enhanced summer monsoon circulation in the Southwestern USA coupled with warmer conditions across most of western North America (Thompson and Anderson, 2000).

Some data indicate considerable temporal variability of the Holocene climate in the southwestern USA (Dean, 1996; Polyak and Asmerom, 2001). For example, uranium-thorium dating of stalagmites from Carlsbad Caverns and Hidden Cave in the Guadalupe Mountains by Polyak and Asmerom (2001) revealed signs of strong climate variability during the second half of the Holocene in New Mexico. According to these authors climate conditions around 4000 years BP comparable with or slightly wetter than the present climate, dominated the southwestern deserts until 3000 years ago. A significant period of increased moisture occurred between 3000 and 1700 years BP. The greater-than-present wetness persisted until about 800 years ago. Afterward, conditions became as dry as or drier than present-day conditions (Polyak and Asmerom, 2001). Two short very dry episodes, between 1130 and 1180 AD, and again between 1275 and 1300 AD, identified by Dean (1996) based on tree-ring studies and archaeological data throughout the southwest, suggest considerable climatic instability in this region during the second half of the Holocene, when very wet episodes were immediately followed by very dry intervals.

III Uncertainties of current trends and predicted scenarios

Arid climates exhibit significant variability at various temporal scales. While the palaeodata provide us with valuable information about the global and regional mechanisms of climate change during relatively long time intervals (centuries and millennia) they do not usually capture the pattern of finer fluctuations of temperature and precipitation. Because interannual climate variations in arid zones are usually very significant and the period of instrumental meteorological observations is still relatively short, interpretation of the recent climate is particularly challenging.

1 Saharo-Arabian deserts

With respect to temperature, by the year 2050 land areas may warm by as much as 1.6°C over the Sahara and the Arabian peninsula (IPCC, 2001). It is expected that the coastal regions will warm more slowly than the continental interior. Annual temperature trends for this and other regions of study are expressed as anomalies based on the Global Historical Climate Network data set (Peterson and Vose, 1997) and shown in Figure 2. The historical climate data show that there has been a general warming trend in all four arid regions of this study (southwest USA, Turanian central Asia, arid Australia and the Saharo-Arabian region) since the beginning of the twentieth century, but with considerable fluctuation along the way.

However, some measurements taken in the southern and eastern Mediterranean do not show warming trends (Ben-Gai *et al.*, 1999). A slight but nonsignificant cooling trend of approximately $-0.5°C\ yr^{-1}$ was detected in Northern Africa, the Negev desert and the Arabian peninsula (Nasrallah and Balling, 1996; Kutiel *et al.*, 2000). To explain the cooling trend observed in the eastern Mediterranean basin, Conte *et al.* (1989) hypothesized a spatial Mediterranean Oscillation between the eastern and

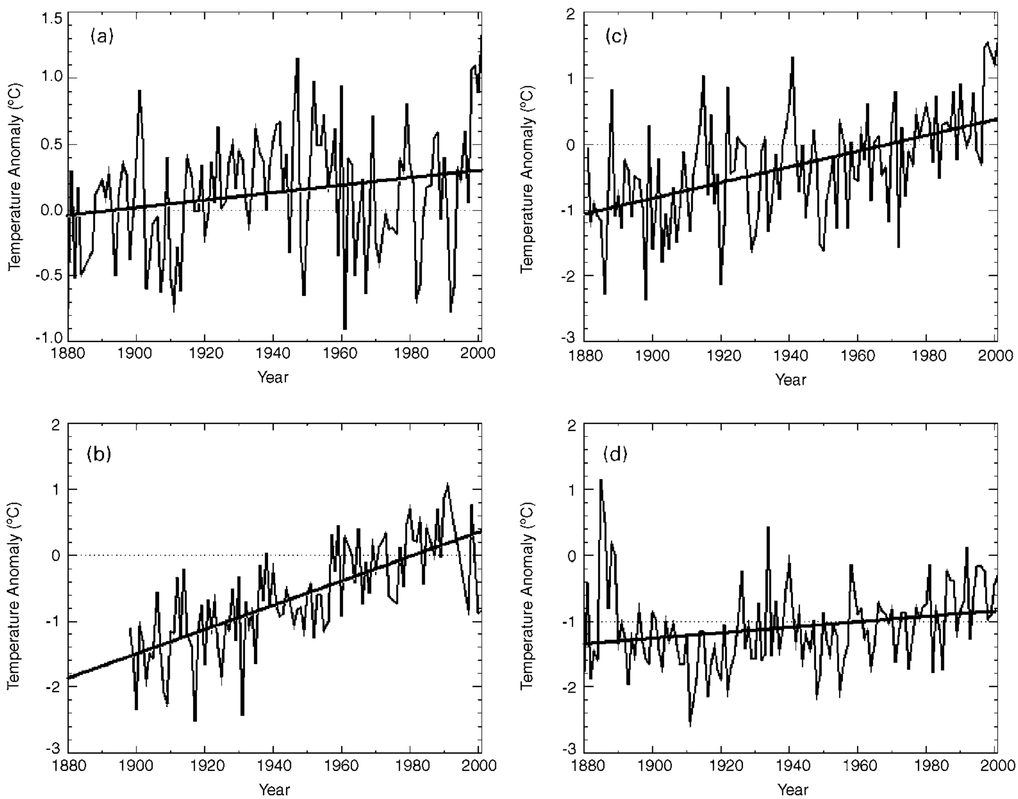

Figure 2 Annual temperature trends since 1900 in (a) the Saharo-Arabian region; (b) central and western Australia; (c) the Turanian deserts of Central Asia; and (d) the southwest USA

western Mediterranean basins (similar to a small-scale El Niño phenomenon). Although such an oscillation has never been proved, several authors found distinctive opposite temperature trends in the western (warming +0.4 C per 100 years) and the eastern sides of the Mediterranean basin (Climate Change Israel, 2000). Rainfall changes in the Saharo-Arabian region projected by most GCMs are relatively modest, at least in relation to present-day rainfall variability. Several models, however, project a reduction of rainfall by approximately 10% over the Horn of Africa and the Arabian peninsula by 2050.

Great uncertainty exists in relation to regional-scale rainfall changes simulated by GCMs (Joubert and Hewitson, 1997; Hulme *et al.*, 1999, Hulme, 2001). While some modelling experiments predict that rainfall will increase by as much as 15% over the 1961–90 average in the Sahel (Joubert and Hewitson, 1997; IPCC, 2001), other studies predict a decrease of precipitation in the same region (Hulme, 2001). It could be expected that local and regional processes would affect precipitation, reducing the reliability of such coarse-resolution models. In addition, the models do not incorporate the effects of dust fluxes in the Saharo-Arabian deserts and urban pollution aerosols along the Mediterranean coast on rain production in this region. Both locally generated aerosols and pollution originating from Europe may affect temperature and precipitation distribution on regional and local scales and greatly affect the spatial accuracy of the GCM scenarios. Because of their failure to incorporate all drivers of the regional climate, the currently available GCM scenarios have relatively low predictive power to project from global to a regional climate change assessment.

2 Australian deserts and semi-deserts

While the Earth has warmed by $0.6 \pm 0.2^{\circ}$C on average since 1900, Australia's continent-average temperature has risen by about 0.7°C from 1910 to 1999 (Commonwealth Scientific and Industrial Research Organization (CSIRO), 2002) The warming has been especially marked since the 1970s and the most recent decade (1989–1998) was the warmest on record (Figure 3b). This recent warming has been greatest in winter and spring. The increase in mean temperature has resulted mostly from increases in minimum (night-time) temperature, with only smaller increases occurring in daytime maxima. Consequently the diurnal temperature range has decreased over the continent by nearly 1°C (Hulme and Sheard, 1999). While Australian rainfall has varied substantially over time and space, there has been no significant continental trend since 1910 (IPCC, 2001; CSIRO, 2002).

GCM simulated ranges of warming for Australia suggest that by 2030 annual average temperatures may be 0.4–2.0°C higher over most of the continent, with potential for greater warming in the northwest. By 2070, annual average temperatures could increase by anything between 1.0°C and 6.0°C according to different GCM scenarios, with spatial variability similar to those for 2030 (CSIRO, 2002). In most scenarios the arid interior of Australia warms more rapidly than the coastal regions. This difference amounted to at least 2°C in the scenarios for 2070–80 (Hulme and Sheard, 1999; CSIRO, 2002). Model results indicate that future increases in daily maximum and minimum temperature will be similar to the changes in average temperature.

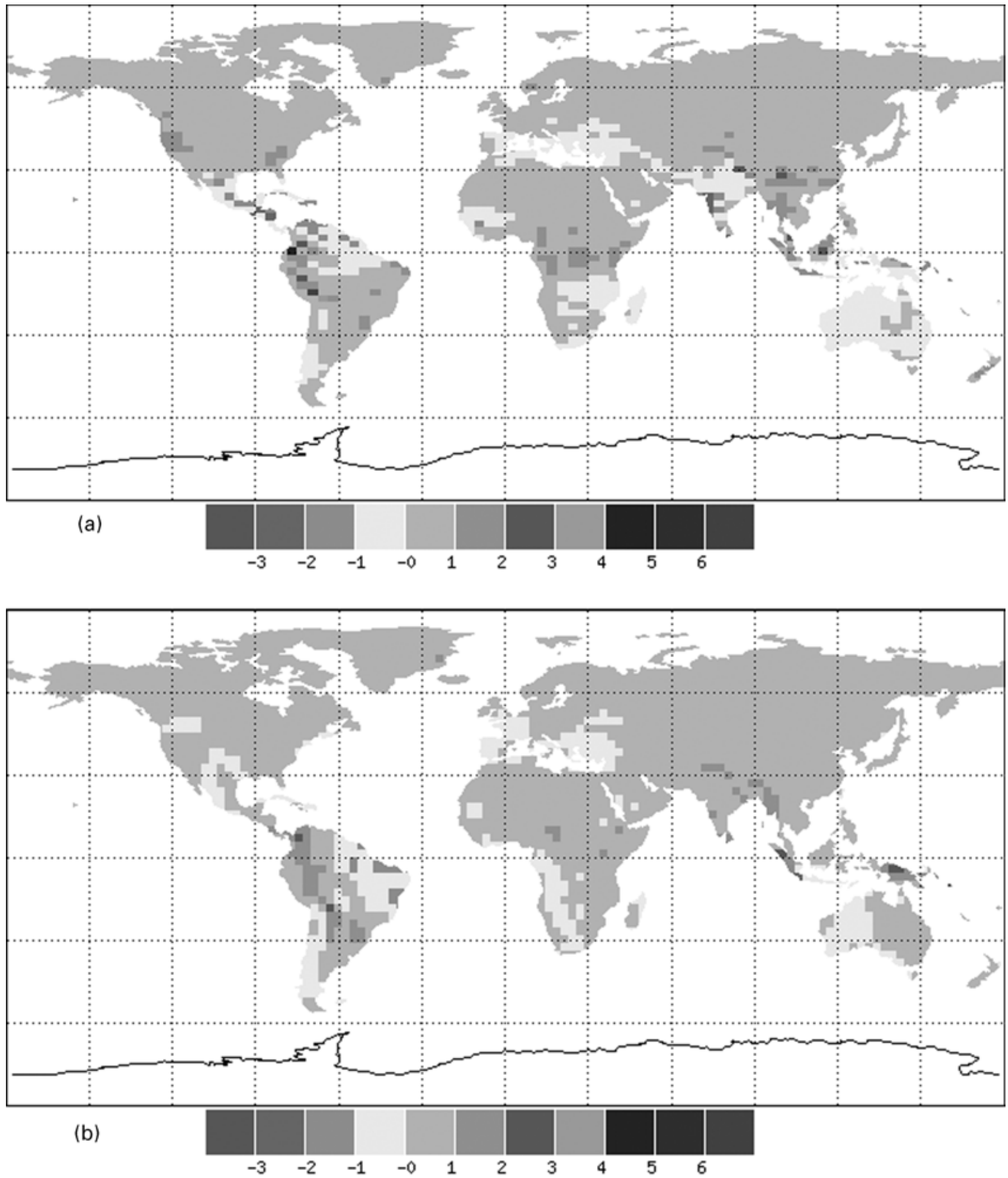

Figure 3 Annual precipitation changes predicted for 2080, relative to 1961–90 by (a) HADCM2 (GG forcing); (b) ECHAM4; (c) CGCM1; (d) CSIRO-MK2 (computed by the IPCC-DDC)

This contrasts with the greater increase in minima rather than maxima observed over Australia in the twentieth century.

Higher temperatures are likely to increase evaporation. CSIRO has calculated projections of change in potential evaporation from eight GCMs. The results show that an increase occurs in all seasons and, annually averaged, ranges from 0% to 8% per degree of global warming over most of Australia.

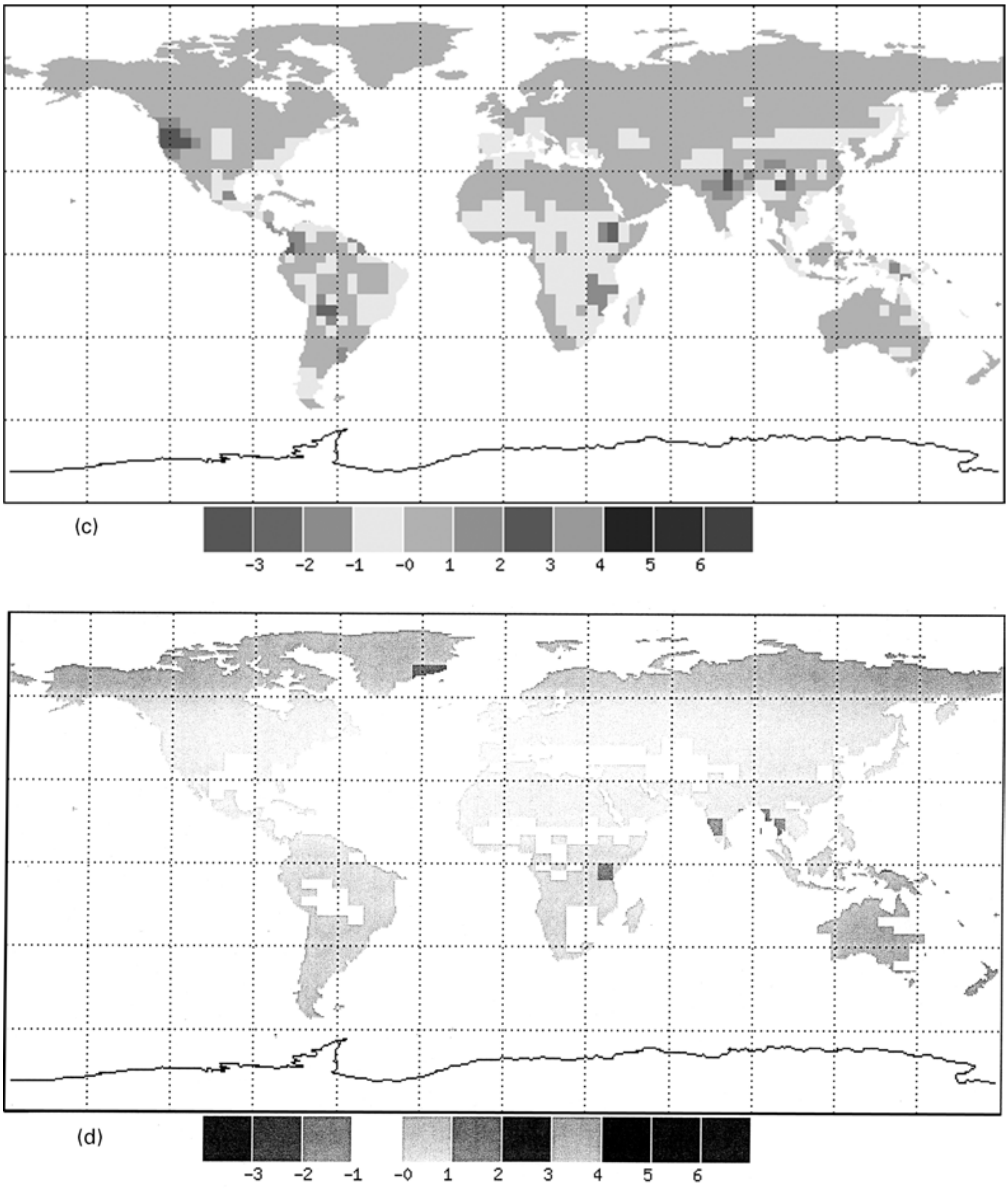

Figure 3 *Continued*

Projected annual average ranges of Australian rainfall tend towards a decrease in the southwest (−20% to +5% by 2030 and −6% to +10% by 2070). In some other areas, including much of eastern Australia, projected ranges are −10% to +10% by 2030 and −35% to +35% by 2070 (CSIRO, 2002) and so represent little change from current conditions.

The difference between potential evaporation and rainfall gives the net atmospheric moisture balance. When the simulated increases in potential evaporation are considered in combination with simulated rainfall change, the overall pattern shows decreases in moisture balance throughout the entire Australian continent. Average decreases in annual water balance range from approximately 40 to 120 mm per degree of global warming and mean greater moisture stress for Australia than any other continent.

3 The Turanian deserts

Records of annual temperature during the period 1900–96 show a 1–2°C per century increase for the arid zone of Central Asia (Uzbekistan, 1999; IPCC, 2001). There was no discernible trend in annual precipitation during the past century. Climate models predict that the temperature in the region will increase by 1–2°C by 2030–50, with the greatest increase in winter. Precipitation projections vary from one model to another (see Table 3) but are unlikely to be significant. Because of the projected increase in temperatures, higher evaporation is expected in Central Asia. Soil moisture is projected to decrease in most parts of this region, which may lead to increased areas of soil degradation (IPCC, 2001). Projected changes in the aridity index for different model runs show no consistent trend (IPCC, 1996, Working Group I, Chapter 14). Some models project greater aridity in the future and some lower. The lack of data specifically from arid and semi-arid regions contributes to this uncertainty.

Vegetation models project little change for most arid vegetation types of the central Asian republics. However, impacts may be greater in the marginal semi-arid zones, such as the grasslands of northern Kazakhstan and the steppe, shrubland and forest ecosystems of eastern Uzbekistan, Tadjikistan and Kyrgyzstan, where even insignificant changes of climate and vegetation under elevated CO_2 levels might favour an increase of human pressure and so contribute to land degradation.

As mentioned earlier, global climate change scenarios do not incorporate the regional controls on climate. Regional climate changes caused by the degradation of the Aral Sea and extensive redirection of water resources to irrigated agriculture in this region is an illustrative example of such human-induced processes, whose impact on regional climate is very significant but not yet fully understood and is not taken into account by the models. Regional weather records show a significant increase in summer and annual air temperature and a decrease in winter temperatures in the vicinity of the Aral Sea. The reduction of the sea surface area has also caused a significant decrease of precipitation in this region since the 1960s and saline dust from the exposed lake bed has been implicated in climate and vegetation change, as well as in health problems and economic disaster (Glantz, 1999).

4 The US deserts and semi-deserts

The arid lands of the western USA experience great spatial variability caused by the very complex topography of this region. While precipitation of the southwestern deserts and semi-deserts is controlled by summer monsoons, highly variable winter precipitation prevails throughout the rest of this region. Historically, the region has experienced exceptionally wet and dry periods. Since 1900, temperatures in the western USA have risen by 1–3°C (US Global Change Research Program, 2001). The region has generally experienced significant increases in precipitation, with increases in some areas greater than 50%. However, a few areas such as Arizona have become drier and have experienced more droughts. The length of the snow season decreased by 16 days from 1951 to 1996 in California and Nevada. Extreme precipitation events have increased (US Global Change Research Program, 2001; IPCC, 2001).

The Hadley Center and Canadian Climate Center GCM scenarios used in the US National Assessment project (2001), predict annual average temperature increases of 2°C by the 2030s and 4.5–6°C by the 2090s. The models project increased precipitation during winter, especially over the mountain ranges of California, where runoff is projected to double by the 2090s. In these climate scenarios, some areas of the Rocky Mountains are projected to get drier. Both the Hadley Center model and Canadian Climate Center model project more extreme wet and dry years. Because of uncertainties about regional precipitation, the possibility of a drier climate should be also considered.

Annual temperature and precipitation changes in four selected arid regions computed by HADCM2, ECHAM4, GFDL-R15, CGCM1, CSIRO-Mk2 GCMs available from the IPCC Distribution Centre, are summarized in Table 2. Note that the HadCM2 projections with greenhouse gas and anthropogenic aerosol forcing (GS) differ slightly from the HADCM2 scenarios with greenhouse forcing only (GG). While all the five models predict the temperature increase in all four arid regions, the precipitation scenario seems to be less consistent.

IV Can the past help to predict the future?

1 Challenges of climate modelling in arid zones

It is generally accepted that physically based computer modelling offers the most effective means of answering questions relating to the prediction of future global climate change and of the potential impacts of climate change (Peterson and Vose, 1997; IPCC, 2001; McGuffie and Henderson-Sellers, 2001). Fully coupled global circulation models (GCMs), simple models and models of intermediate complexity have been used during the past decades to explore the trends of global climatic change. The GCMs are used for calculation of the complex full three-dimensional character of the climate comprising the global atmosphere (AGCMs) and/or the ocean (AOGCMs or CGCM) and possibly other components, such as changing biomes (AOBGCMs) and biogeochemical changes (McGuffie and Henderson-Sellers, 2001). Simple models allow one to explore the potential sensitivity of the climate to a particular process over a wide range of parameters (Kattenberg *et al.*, 1996).

Recently, significant advances have occurred in the development of Earth System Models of Intermediate Complexity (EMIC), which are designed to bridge the gap between the three-dimensional comprehensive models and simple models. The main characteristic of EMICs is that they describe most of the processes implicit in comprehensive models, albeit in a more parameterized form (IPCC, 2001; Kageyama, 2001). They also simulate explicitly the interactions among several components of the climate system, including biogeochemical cycles. On the other hand, EMICs are sufficiently computationally efficient to allow for long-term climate simulations over several tens of thousands of years or a broad range of sensitivity experiments over several millennia. Examples of such models are CLIMBER-2, used for transient

Table 2 Annual temperature and precipitation changes in four selected arid regions predicted by five GCMs

GCM	Institution	Forcing type	Saharo-Arabian 2020	Saharo-Arabian 2050	Saharo-Arabian 2080	Australian 2020	Australian 2050	Australian 2080	Turanian 2020	Turanian 2050	Turanian 2080	American West 2020	American West 2050	American West 2080
(a) Annual temperature changes (range, °C)														
HadCM2 (I92a scenario)	Hadley Center	GG	+1 to +3	+3 to +4	+4 to +6	+1 to +2	+2 to +4	+5 to +6	+1 to +2	+1 to +3	+4 to +5	+1 to +2.5	+2 to +4	+4 to +5
		GS	+1 to +2	+2 to +3	+3 to +5	+1 to +2	+2 to +3	+4 to +5	+1 to +2	+1 to +2.5	+3 to +4	+1 to +2	+2 to +3	+3 to +4
ECHAM4	Max Planck Institute	GG	+1 to +2	+2 to +4	+4 to +5	+1 to +2	+3 to +4	+4 to +5.5	+1 to +2.5	+3 to +4	+4 to +5.5	+1 to +2.5	+2 to +4	+4 to +5.5
GFDL-R15	Geophysical Fluid Dynamics Laboratory	GG	+1 to +3	ND	ND	+2 to +3	ND	ND	+2 to +3	ND	ND	+1 to +3	ND	ND
CGCM1	Canadian Climate Center	GG	+2 to +3	+3 to +5.5	≥+5	+2 to +3	+3 to +5	≥+5	+2 to +4	+3 to +5	≥+5	+2 to +3	+3 to +5	≥+5
CSIRO-Mk2	Commonwealth Scientific and Industrial Research Organization	GG	+1 to +2	+2 to +3.5	+3 to +5	+1 to +2	+2 to +3	+3 to +4	+1 to +2	+2 to +3.5	+3 to +5	+1 to +3	+2 to +3.5	+3 to +5
(b) Annual precipitation changes (range, mm day^{-1})														
HadCM2 (I92a scenario)	Hadley Center	GG	−1 to −1	−1 to 1	−1 to 1	−1 to 1	0 to 1	−1 to 0	0 to 1	0 to 1	0 to 1	0 to 1	0 to 2	1 to 2
		GS	0 to 1	0 to 1	0 to 1	0 to 1	−1 to 0	−1 to 0	0 to 1	0 to 1	0 to 1	0 to 1	0 to 1	0 to 2
ECHAM4	Max Planck Institute	GG	0 to 1	0 to 1	0 to 1	0 to 1	−1 to 1	−1 to 1	0 to 1	0 to 1	0 to 1	0 to 1	0 to 1	0 to 1
GFDL-R15	Geophysical Fluid Dynamics Laboratory	GG	0 to 1	ND	ND	−1 to 1	ND	ND	0 to 1	ND	ND	0 to 1	ND	ND
CGCM1	Canadian Climate Center	GG	0 to 1	−1 to 1	−1 to 1	0 to 1	−1 to 1	−1 to 1	0 to 1	−1 to 1	−1 to 1	0 to 1	0 to 2	1 to 3
CSIRO-Mk2	Commonwealth Scientific and Industrial Research Organization	GG	0 to 1	−1 to 1	−1 to 1	0 to 1	−1 to 1	−1 to 1	0 to 1	−1 to 1	−1 to 1	0 to 1	0 to 1	0 to 1

ND, no data.

simulations of the Holocene climate in the Sahara (Claussen *et al.*, 1999, 2003; Kubatzki *et al.*, 2000), or the Quasi-Equilibrium Tropical Circulation Model (QETCM) used by Zeng and Neelin (2000) to explore climate variability in Africa. EMICs are simpler and cheaper than GCMs, which makes them more appropriate for longer timescale studies.

Despite the great progress in climate modelling during the last decades, there is still considerable disagreement between different models (Table 2; Figure 3). Although such disagreements and uncertainties are a feature of climate change modelling in many climate zones, the scenarios projected for arid zones under global warming are especially variable (Joubert and Hewitson, 1997; Hulme *et al.*, 1999, 2001). Perhaps the main reason for the high uncertainty of arid climate modelling is the extreme natural variability (both temporal and spatial) of the desert climate. The very existence of arid ecosystems is related to their temporal and spatial environmental variability. Arid climates exhibit different degrees of interannual, interdecadal, multidecadal, and interseasonal variations of climate (especially precipitation). Together with vegetation cover characteristics (such as albedo (reflectivity), leaf-area index or LAI (area of leaves per unit ground area), roughness length and canopy density and height) they represent the major challenge for climate forecasting and modelling in arid zones. This temporal land-cover variability may in turn, be a product of totally different mechanisms of vegetation–climate feedback controlling climate during the wet and dry time intervals in the present-day climate, just as in the past. It is well known from climate model studies on seasonal to annual timescales that natural climate variability has a great impact on the outcome of a given model scenario (McGuffie and Henderson-Sellers, 2001; Hulme *et al.*, 2001; Renssen *et al.*, 2002). Atmospheric dynamics are known to be very sensitive to natural climate variability on relatively short timescales. The effect of the short time variability on a longer (decadal to millennial) timescale has not yet been thoroughly studied in ensemble experiments with fully coupled AOGCMs, as such experiments are extremely expensive (Renssen *et al.*, 2001).

Although temporal variability of precipitation in arid zones is emphasized by many studies, spatial variability is equally important as the driving variable for ecosystem processes (Whitford, 2002). Precipitation in arid environments is governed by topography. The extensive arid regions of the Great Basin of North America and the Turanian deserts of Central Asia are classic examples of deserts resulting from orographic effects on the flow of air masses. The mechanisms of mid-latitude aridity, however, are not limited to 'rainshadow effects' of the mountains but are largely the effects of mountains on the polar jet stream. The interaction of mountain ranges with the polar jet stream determines regions of frequent passage of intertropical disturbances (Manabe and Broccoli, 1990; Whitford, 2002).

Both frontal and convectional precipitation in arid zones exhibit a strong cellular pattern with a great variability of cell size. Because of the very low spatial correlation of intense rainfall in deserts, even areas experiencing the same storm can receive very different amounts of rainfall. Because the climate model grid sizes are much larger than convective elements in the atmosphere, the precipitation pattern is still poorly represented not only in the GCMs but also in the regional climate models. The spatial scale of existing models and downscaling techniques are still insufficient to understand climatic data in their regional and local context. Regional climate models or nested models will help improve our understanding of the regional-scale phenomena that are important for climate trends in arid zones. However, RCM experiments on arid zones run in climate change mode are very few and usually are of short duration, and therefore are not able to resolve decadal-scale climate variability (Hulme, 2001; McGuffie and Henderson-Sellers, 2001; Hulme *et al.*, 2001).

There is strong agreement among climatologists that the climate transition from LGM conditions to the Holocene was triggered by changes in seasonal sunlight distribution caused by oscillations in the Earth's orbit and the tilt of the Earth's axis (Berger, 1978; Kutzbach and Guetter, 1986). The Earth's orbital parameters are believed to be the main factor controlling the intensity of monsoon intensity in the Northern Hemisphere. If the changes in desert climates that occurred during the Pleistocene and Holocene were due only to insolation effects, there is little in the Holocene optimum climate that could act as an analogue of the greenhouse-forced climate change. At least the physical differences of climate change forcings imply that one may expect very different responses of arid climates to a future global trend compared with the Holocene climate in terms of the frequency, rapidity and amplitude of such changes. However, it is important to consider many other global and regional factors that could affect climate variability of arid zones in the past just as today.

Both theoretical considerations and numerical models have shown the significant sensitivity of the climate of arid regions to vegetation distribution (Claussen *et al.*, 1999; Wyputta and McAvaney, 2001; Wang and Eltahir, 2000) and also the possibility that this resulted in globally significant changes of reservoirs of organic carbon in vegetation and soils in these regions, affecting atmospheric CO_2 (Peng *et al.*, 1995; Petit-Maire *et al.*, 1995; Lioubimtseva *et al.*, 1996; 1998; Lioubimtseva, 1999). Recent modelling experiments suggest that relatively abrupt climate changes between the LGM and the Holocene, and later from the mid-Holocene to present-day climate, were ultimately caused by seasonal sunlight variations but strongly amplified by atmosphere–vegetation feedbacks (Claussen *et al.*, 1999, 2003; Diffenbaugh and Sloan, 2002).

It was demonstrated in several modelling studies (Wang and Eltahir, 2000; Zeng and Neelin, 2000; Claussen *et al.*, 2003) that vegetation plays a prominent role in the energy, moisture and carbon exchange between the land surface in arid and semi-arid zones and the atmosphere. When a vegetation–climate system is perturbed by an external or internal factor (such as insolation changes in the past or human-induced changes today), the system can respond in three qualitatively different ways: a negative feedback leading to a full recovery, a negative feedback leading to a partial recovery or a positive feedback leading to a perturbation enhancement. Therefore the vegetation–climate system can have multiple equilibrium states coexisting under the same forcing (Wang and Etahir, 2000).

Ground-cover parameters can significantly alter the modelled climate. For example, the recent modelling experiments by Claussen *et al.*, (1999, 2003), showed that a positive feedback between vegetation and precipitation is critical for understanding the rapid expansion of savanna vegetation into the Sahara in the early and mid-Holocene (9000–6000 years Bp), both for simulations with orbital forcing and greenhouse forcing. This, however, does not mean that there is a direct analogue between the reduction of desert zones during the Holocene climate optimum and the predicted greenhouse-gas-induced climatic change.

On the other hand, palaeoclimate modelling experiments forced only by orbital parameters and that do not take into account biophysical and biochemical feedbacks of the Holocene vegetation cover, tend to produce a simplistic picture of climate change in the past. For example, in the PIMP experiments (Palaeoclimate Modelling Intercomparison Project) by Prentice *et al.* (1998) the AGCM outputs were used by the biogeography model (BIOME6000) to enable comparison between the modelled biome distribution in the Holocene and available palaeodata. Significant disagreement between the palaeoreconstructions and palaeomodelling results may be caused by the failure to incorporate complex climate–vegetation feedbacks into the AGCM.

Another factor that needs to be considered in climate simulations of arid zones is the role of changes of the global and regional reservoirs of organic carbon in vegetation and soils caused by land use and landcover changes in these regions. Conversion of semi-deserts and dry steppes into croplands and pastures often leads to significant carbon losses in these ecosystems. It is estimated (Dregne and Tchou, 1992) that of approximately 5160 Mha of drylands (which exclude hyper-arid deserts), 69% has been degraded, mainly by the loss of vegetation cover but also by soil degradation, mainly erosion. Lal (2001) assumed that land degradation in arid regions could lead to an average reduction in the soil organic carbon pool of 8–12 carbon ha^{-1}. Unfortunately, there have been no comprehensive assessments of the amount of carbon lost strictly through degradation of arid zones. The level of arid land degradation is highest in North America (85%) and lowest in Australia (55%). According to Keller and Golstein (1998), the potential land available for restoration and carbon storage in arid zones is on the order of 3500 Mha. That, along with possible carbon sequestration in other climate zones, may have a significant impact on the global carbon budget.

2 Possible implications of climate change for arid ecosystems

In addition to its effect on climate, an increased atmospheric CO_2 concentration has direct and relatively immediate effects on two important physiological processes in plants – it increases photosynthetic rate but decreases stomatal opening and therefore the rate at which plants lose water. Combination of these two factors implies a significant increase of water efficiency (the ratio of carbon gain per unit water loss) in desert vegetation as a result of elevated atmospheric CO_2 (Huxman and Smith, 2001). Theoretical considerations and biogeography models (Mellilo *et al.*, 1993; Neilson and Drapek, 1998, Woodward *et al.*, 1998; Cramer *et al.*, 2001) predict relatively strong responses of arid ecosystems to global climatic change. Elevated CO_2 concentrations can, among other effects, enhance productivity and increase the water use efficiency (WUE, carbon fixed per unit water transpired) of the vegetation, thereby reducing the sensitivity of the vegetation to drought stress (Neilson and Drapek, 1998; Bachelet *et al.*, 2001; Cramer *et al.*, 2001). The results of simulations by the leading dynamic global vegetation models for arid lands (Table 3) suggest that the direct CO_2 effect may be the key factor of ecosystem changes in these zones.

Simulation of changes in arid lands under various GCM scenarios vary strongly depending on whether or not the direct effects of elevated CO_2 have been incorporated. The water-controlled boundaries may exhibit any direction of change, depending on the interaction of several variables, such as the annual and seasonal changes in

Table 3 MAPSS and BIOME3 scenarios of changes in arid zones of the world

Scenarios simulated by the MAPSS and BIOME3 for arid lands	With CO_2 effect (%)		Without CO_2 effect (%)
	Older (first assessment) scenarios	Newer (second assessment) scenarios	Newer (second assessment) scenarios
Percentage of current biome area in future under various GCM scenarios	71–72	59–78	83–120
Percentage of current biome area, which could undergo a loss of LAI (i.e., biomass decrease) owing to global warming	8–12	0–13	0–29
Percentage of current biome area, which could undergo a gain of LAI (i.e., biomass increase) owing to global warming	51–57	53–80	23–60
Percentage of current biome area, which could undergo a loss of annual runoff owing to global warming	24–26	1–20	2–20
Percentage of current biome area, which could undergo a gain of annual runoff owing to global warming	7–25	4–15	3–15

Source: based on Regional Impacts of Climate Change, IPCC 2001 (Annex C, tables C-1, C-2, C-3, C-4, and C-5).

temperature and precipitation and the direct physiological effects of CO_2 on plant productivity and water use efficiency.

The continental- and regional-scale modelling experiments conducted in the US (Neilson and Drapek, 1998; Bachelet *et al.*, 2001) showed a 60% or greater reduction in the area of deserts under Hadley Centre (HadCM2) and Canadian Climate Centre (CGCM1) scenarios. Under both the HadCM2 and CGCM1 scenarios, static biogeogaphy models, such as LPJ, BIOME3 (Haxeltine and Prentice, 1996), MAPSS (Neilson and Drapek, 1998) as well as a more recent dynamic biogeography model, such as MC1 (Bachelet *et al.*, 2001) suggest an increase in plant growth, a reduction in desert areas and a shift toward more woodlands and forests in many parts of western USA. The biogeography model simulations also indicate up to a 200% increase in leaf area index in the US deserts and semi-deserts and a northern expansion and migration of arid-land species into the Great Basin region. These modelling scenarios are often interpreted by policy makers and groups opposed to the greenhouse gas emission regulations as an indication that global warming will likely enhance the agricultural potential of the arid zones. However, the modelling experiment also shows that a further rise in temperature may cause the return of and further intensification of aridity (Neilson and Drapek, 1998; Bachelet *et al.*, 2001). In addition, the possibility of a less positive CO_2 fertilization effect than is assumed in the models, an increase in fires and persistence of other stresses such as air pollution, are important sources of uncertainty. Given the complexity of mechanisms of ecosystem responses to global and regional climate change, it seems appropriate to treat model scenarios with some caution. The short-term responses of ecosystems to greenhouse-forced climate change may in turn create a totally new regional pattern of vegetation–climate relationships, very different from those in the low CO_2 Holocene world. There is a clear need for a further research requiring a synergy between the modelling efforts and field experiments studying ecological responses of arid ecosystems to global change.

A significant amount of experimental work has been carried out on CO_2 enrichment effects in deserts and desert margins during recent years. For example, the Nevada desert Free-Air CO_2 Enrichment Facility (Hamerlynck *et al.*, 2000; Huxman *et al.*, 2000; Smith *et al.*, 2000), which has been in operation since 1997, is attempting to predict the possible complex ecological and biogeochemical changes in semi-desert ecosystems caused by increasing atmospheric CO_2. Hamerlynck *et al.* (2000) determined that a 52% increase in the air's CO_2 content increased short-term photosynthetic rates in creosote by 100% and 80% during the wet and dry seasons, respectively. In addition, because elevated CO_2 did not affect rates of stomatal conductance, the water-use efficiency of this species was similarly enhanced by 100% and 80%, respectively. However, actual growth rates, rather than short-term photosynthetic efficiency and water use, may not change much under raised CO_2. Although some seasonal differences were observed between root growth of *Larrea* under ambient and raised CO_2, the year-averaged root growth rate of both *Larrea* and *Ambrosia* was not significantly different between the treatments (Huxman *et al.*, 2000). Early results indicate that semi-desert plants respond especially strongly to raised CO_2 during the occasional wet years that correspond to El Niño events. There is greater year-to-year variation in production cycles at elevated CO_2, suggesting that this system may become even more episodic, and thus in this sense more 'desert-like' in a future high-CO_2 world (Huxman and Smith, 2001).

Also of importance is the fact that non-native invasive grasses may respond to CO_2 such that they are far more productive than native plants during wet years (Smith *et al.*, 2000). *Bromus* invasions in the Great Basin region are known to increase the frequency of fires from a 75–100 year cycle to a 4–7 year cycle. These fires are far more intense than those in native vegetation and usually result in a loss of native shrubs. A change from shrubs to grasses under raised CO_2 would have a dramatic effect on desert water cycles and wildlife habitat, as well as on socioeconomic factors. Based on the very limited number of free-air and chamber CO_2 fertilization experiments on semi-arid systems, we should probably not expect a large response by the vegetation as a whole (as predicted by the models) but by certain individual species, which apparently arbitrarily show a very large response when most others barely respond at all. It should be also taken into account that overgrazing often results in vegetation species composition (towards C4-dominated ecosystems) that may considerably reduce the predicted response of arid ecosystems to the increasing CO_2. Also, it is important to bear in mind that although the FACE experiments provide very important information that might improve our understanding of the future changes in arid ecosystems caused by direct CO_2 effects, they do not incorporate other aspects of climate change, such as seasonal and annual temperature and precipitation changes, that are likely to accompany the changes in atmospheric CO_2 levels.

V Conclusions

So, is the past a key for the future as is often assumed? The answer could be yes and no. Yes, because the palaeoreconstructions can improve our understanding of the role of vegetation–climate feedbacks caused by ground-cover biophysical changes and probably also by changes in global carbon sources and sinks. No, because the physical forcings of climate change in the Holocene and in the predicted greenhouse scenarios are completely different. They may most likely result not only in totally different mechanisms and patterns of climate change but also very different vegetation responses related to plant physiology.

Arid climates exhibit different degrees of temporal (interannual, interdecadal, multidecadal and interseasonal) and spatial variations of climate (especially precipitation) and vegetation cover characteristics. These represent the major challenge for climate forecasting and modelling. However, there has been a general warming trend in all four arid regions of this study (southwest USA, Turanian central Asia, arid Australia and the Saharo-Arabian region) since the beginning of the twentieth century, but with considerable fluctuation along the way. There is no evidence of a strong overall trend towards drier or wetter conditions during the same period for any completely arid region, although there have been considerable decadal timescale fluctuations. Such fluctuations over the past century seem to be comparable by their amplitude and frequency with the earlier climate variability of the second half of the Holocene. These short-term trends also show significant spatial variability caused by regional topography and other factors. For example, while the southwestern USA has generally experienced significant increases in precipitation compared to the past century, with increases in some mountainous areas greater than 50%, some parts of this region, such as Arizona, have become drier and have experienced more droughts.

Precipitation in desert areas is often predicted to increase generally under 'greenhouse' warming during the next century. Despite the great progress in global climate modelling the GCMs give very variable results, with large spatial differences in the areas forecast to give higher or lower precipitation. The precipitation trends seen over the last century do not always agree with older GCM scenarios and there is also a considerable disagreement among different GCMs regarding regional-scale rainfall changes in arid zones. The lack of integration of such factors as dust and aerosol fluxes, hydrological and geomorphological responses to global change, biophysical and biochemical feedbacks caused by land cover, as well as numerous regional and local factors not taken into account by the models, could cause such disagreements. Although the spatial resolution of existing models and downscaling techniques has improved during the last decade their spatial scale is still insufficient to understand climatic data in their regional and local context. Regional climate model experiments on arid zones run in climate change mode are very few and are not able to resolve decadal- or annual-scale climate variability.

Projections based on biogeography models suggest considerable changes in desert and semi-desert vegetation as a result of a combination of greenhouse-related climate change and direct physiological CO_2 effects on vegetation, such as changes in photosynthesis and water use efficiency over the coming century. However, direct-CO_2 experiments with desert vegetation show considerable complexity in responses, with the same species responding completely differently in different experiments, and some species responding far more strongly than others. The recent FACE experiments can provide relatively limited information about the ecosystem responses to climate change because they study only the direct physiological responses of vegetation to the increased CO_2 levels. However, they do not take into account other aspects of climate change, such as temperature and precipitation changes.

Consideration of the past reveals that desert environments indeed have the capacity to vary dramatically over time. Palaeodata provide important information about spatial patterns of change in the world's deserts in the past, that can significantly improve understanding of the global and regional controls on climate change in arid regions. Not only were there dramatic changes between glacial and interglacial periods, but there are also signs of the Holocene variability at finer temporal scales in all the world's major desert regions. Such strong temporal variability has been noted in Saharan moist and wet phases, the North American and Australian dune-building episodes and the desert margin areas of Central Asia. The Holocene climate variability could be an indication of the possibility of multiple equilibrium states of the climate–vegetation system in arid zones coexisting under the same insolation and greenhouse-level conditions during the Holocene. However, because the mechanisms of palaeoclimatic variability were different from those caused by the current greenhouse warming, any palaeoanalogies should be treated with a great caution.

Research priorities in the area of variability of desert climates include:

1. modelling the complexities of biophysical and biochemical climate–vegetation feedbacks, considering that the climate–vegetation system can have multiple equilibrium states coexisting under the same external or internal forcings (such as CO_2 increase, insolation change or land-cover change);
2. experimental work on direct CO_2 effects on desert and semi-desert ecosystems, incorporating a wider range of the world's arid regions (most work has so far been in the USA), and concentrating on free-air studies of relatively undisturbed systems rather than chamber experiments. The FACE experiments would provide more realistic results if they could take into account not only increased CO_2 levels but also temperature and precipitation changes caused by global climate change;
3. identification of both internal and external factors of climate changes in arid environments over the past tens of thousands of years, as well as the scale and variability, as this may give clues to the responsiveness of deserts to future climate change and their inherent tendency to undergo sudden climate changes on a scale not experienced during the period of instrumental observations.

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