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A global perspective on African climate

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Abstract We describe the global climate system context in which to interpret African environmental change to support planning and implementation of policymaking action at national, regional and continental scales, and to inform the debate between proponents of mitigation v. adaptation strategies in the face of climate change. We review recent advances and current challenges in African climate research and exploit our physical understanding of variability and trends to shape our outlook on future climate change. We classify the various mechanisms that have been proposed as relevant for understanding variations in African rainfall, emphasizing a "tropospheric stabilization" mechanism that is of importance on interannual time scales as well as for the future response to warming oceans. Two patterns stand out in our analysis of twentieth century rainfall variability: a drying of the monsoon regions, related to warming of the tropical oceans, and variability related to the El Niño–Southern Oscillation. The latest generation of climate models partly captures this recent continent-wide drying trend, attributing it to the combination of anthropogenic emissions of aerosols and greenhouse gases, the relative contribution

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of which is difficult to quantify with the existing model archive. The same climate models fail to reach a robust agreement regarding the twenty-first century outlook for African rainfall, in a future with increasing greenhouse gases and decreasing aerosol loadings. Such uncertainty underscores current limitations in our understanding of the global climate system that it is necessary to overcome if science is to support Africa in meeting its development goals.

1 Introduction

This paper discusses the global-scale mechanisms that affect African rainfall. The goal is to place African environmental change in the context of the global climate change debate, so that climate science can play an appropriate role in discussions of African development. Motivation comes from recent advances that point to global climate as a primary agent of African environmental change—in particular, research confirming the dominant role of the global tropical oceans in effecting late twentieth century drought in the Sahel. These advances openly challenge the historical focus on the local nature of causes of environmental degradation and their remedies. They point to the need to create a synergy across scales, between global and local knowledge, to fully explain past environmental change and to guide in the definition of future management strategies (Reynolds et al. 2007).

Advances in the physical understanding of African environmental change that point to global climate as an important agent include: (1) remotely sensed observations of vegetation cover; (2) modeling of land cover response to land use and to precipitation, and (3) climate modeling of the response of African rainfall to the observed distribution of sea surface temperature (SST). Continuous remotely sensed observations of land cover started in the late 1970s. They exposed the ability of vegetation to recover from drought at both seasonal and interannual time-scales (Tucker et al. 1991; Helldén 1991; Anyamba and Eastman 1996; Prince et al. 1998; Nicholson et al. 1998; Eklundh and Olsson 2003; Herrmann et al. 2005). Modeling of land cover and climatic responses to land use have begun to converge towards the view that land use/land cover changes are of secondary importance for African rainfall variability (Taylor et al. 2002; Hickler et al. 2005). In sharp contrast, climate models forced only by the observed long-term history of SST have reproduced the historical record of Sahel rainfall (Giannini et al. 2003; Lu and Delworth 2005), confirming the role of the global oceans in pacing drought in this region (Folland et al. 1986; Palmer 1986; Rowell et al. 1995), as well as more generally over tropical Africa (Hoerling et al. 2006).

Equally relevant are studies of local interpretation and adaptation to climate variability and change. In regards to the recent "greening" of the Sahel—the observation that the state of vegetation cover in much of the region has considerably improved since the very dry mid-1980s (Herrmann et al. 2005)—the remote sensing work must be reconciled with evidence from local surveys. In recent years local practitioners and farmers have not observed a return to pre-drought conditions in the characteristic development of the rainy season (L'Hôte et al. 2002; West et al. 2007). However, when financially feasible, communities have enthusiastically adopted soil water conservation strategies that have locally improved cereal yields (e.g. Reij et al. 2005) and vegetation cover (Polgreen 2007). So, as scientific understanding comes to the conclusion that it was not mismanagement of land resources that caused the persistence of drought, it could very well be that investment in land management practices, not a return of rainfall to pre-drought levels, are found to have most significantly contributed to the greening.

This study focuses on one external cause of environmental change—climate. We focus on the recent observational record of continental-scale tropical African climate variability and on simulations of future change to address the following questions: Are there current trends? Is it wise to linearly extrapolate them into the future? The tropical African focus means that this paper excludes the northern and southern extremities of the continent, whose Mediterranean climates are strongly influenced by extratropical atmospheric circulations. The focus on seasonal-to-interannual and longer-term climate variability derives from our ultimate goal of projecting future African climate change. Complementing the recent investment in regional climate modeling at African institutions—an investment undertaken with a view to its applicability to adaptation to climate variability and change (e.g. Jenkins et al. 2002, 2005; Tadross et al. 2005; Afiesimana et al. 2006; Pal et al. 2007; http://www.aiaccproject.org/)—the focus here remains on the large scales, from continental to global. Given the gaps in our understanding of the large-scale dynamics of climate that are manifest in our inability to explain the uncertainty in projections of future change, a detailed, downscaled analysis is premature. However, we anticipate that projections on regional, rather than continental, scale, and of other characteristics of rainfall besides the seasonal total (such as its intraseasonal variability, distribution of frequency and intensity of events, and monsoon onset date) will become more useful as our outlook on the future becomes less uncertain. Translating a climate outlook on the continental scale—say, for example, a continent-wide drying—into information useful at the regional scale—say, for example, a postponement of the West African monsoon onset, or a change in duration or frequency of dry spells within the southern African wet season (e.g. Usman and Reason 2004)—remains a challenge that will be met when scientific capacity is integrated with stakeholders' demand for climate information (Washington et al. 2004, 2006; IRI 2006).

In Section 2 we review the climatological seasonal cycle over Africa and introduce relevant governing mechanisms. Section 3 is about climate variability and change during the twentieth century, with emphasis on recent decades, as observed, and as simulated in state-of-the-art coupled ocean-atmosphere models. The emphasis is on continental-scale patterns, such as a multi-decadal drying or the impact of the El Niño–Southern Oscillation (ENSO). Section 4 describes the uncertainties in projections of African climate change and discusses the large-scale mechanisms that we see to be potentially involved. Section 5 offers conclusions and recommendations.

2 Dynamics of tropical rainfall and seasonality of African climate

Deep convection is the physical process which brings about most precipitation in the tropics. It is the result of a hydrodynamic instability in which fluid rises through the depth of the troposphere[1] due to its being lighter than the fluid surrounding it, while

[1]The lowest 15–20 km of the Earth's atmosphere, containing 90% of the atmosphere's mass.

heavier fluid sinks, much like water in a pot heated from below. In the atmosphere, the heating from below is due originally to solar (short-wave) radiation, much of which passes through the atmosphere to heat the surface (whether ocean or land). The surface returns that energy to the atmosphere partly by terrestrial (long-wave, infrared) radiation, but also by turbulent fluxes of sensible heat (that which directly influences temperature) and, more quantitatively important in most tropical regions, latent heat in the form of water vapor. The fact that this vapor can be transported for large distances before releasing its latent heat gives tropical climate dynamics much of its distinctive flavor and complexity.

Deep convection can be thought of as a two-step process: first, surface air is warmed or moistened. The warming itself is never large enough to create the low densities needed to convect, i.e. lift the air parcel, through the depth of the troposphere. Rather, the water vapor in the air must be sufficient so that when lifted and cooled, by small-scale turbulence or a large-scale weather disturbance, enough condensation will occur, and enough latent heat will be released to render the parcel sufficiently buoyant to ascend through the depth of the troposphere. In the process the parcel will return most of its original water vapor to the surface in the form of precipitation. The key step is the initial warming and moistening of the surface air, a transformation which renders it potentially light enough to ascend through the troposphere. Note also that a warmer (and thus lighter) troposphere above the surface will suppress deep convection since the near-surface air must then become lighter still in order to become buoyant.

Since tropical precipitation is a response to instability, one way to perturb the system is to change its stability, either by changing the temperature and humidity of near-surface air, or by changing the temperature of the atmosphere at higher levels. It is a fundamental feature of the tropical atmosphere that it cannot sustain large horizontal temperature gradients above the planetary boundary layer (that is, more than 1–2 km above the surface). Warming generated by heating of one part of the tropical atmosphere is efficiently distributed horizontally throughout the tropics. A warming of the air above the planetary boundary layer over the Indian Ocean for example, once communicated to Africa by atmospheric dynamics, would reduce the instability of a boundary layer parcel, and thus inhibit precipitation (Chiang and Sobel 2002). We refer to this mechanism as *stabilization*.

Another way to perturb the system is to change the supply of moisture flowing into the region of interest. Arguments to this effect tend not to directly address the stability properties of the atmospheric column, but instead treat the moisture budget as central. The idea is that other factors, such as gradients in sea surface temperature (Lindzen and Nigam 1987), or the temperature contrast between land and sea (Webster 1994; Hastenrath 1991), induce changes in the pressure field, and thus in the large-scale flow field, so as to alter the low-level convergence of moisture into the region. Since precipitation in rainy regions is largely fed by this moisture convergence rather than by local evaporation, this leads to changes in rainfall. We label this mechanism *moisture supply*.

We will refer to these mechanisms, stabilization and moisture supply, in our discussion of African climate variability and change in the sections to follow.

Seasons over Africa, as elsewhere, result from interactions between atmosphere, land and ocean as they respond to the annual cycle of insolation—the ultimate driving force behind the alternation of wet and dry, warm and cold seasons. The

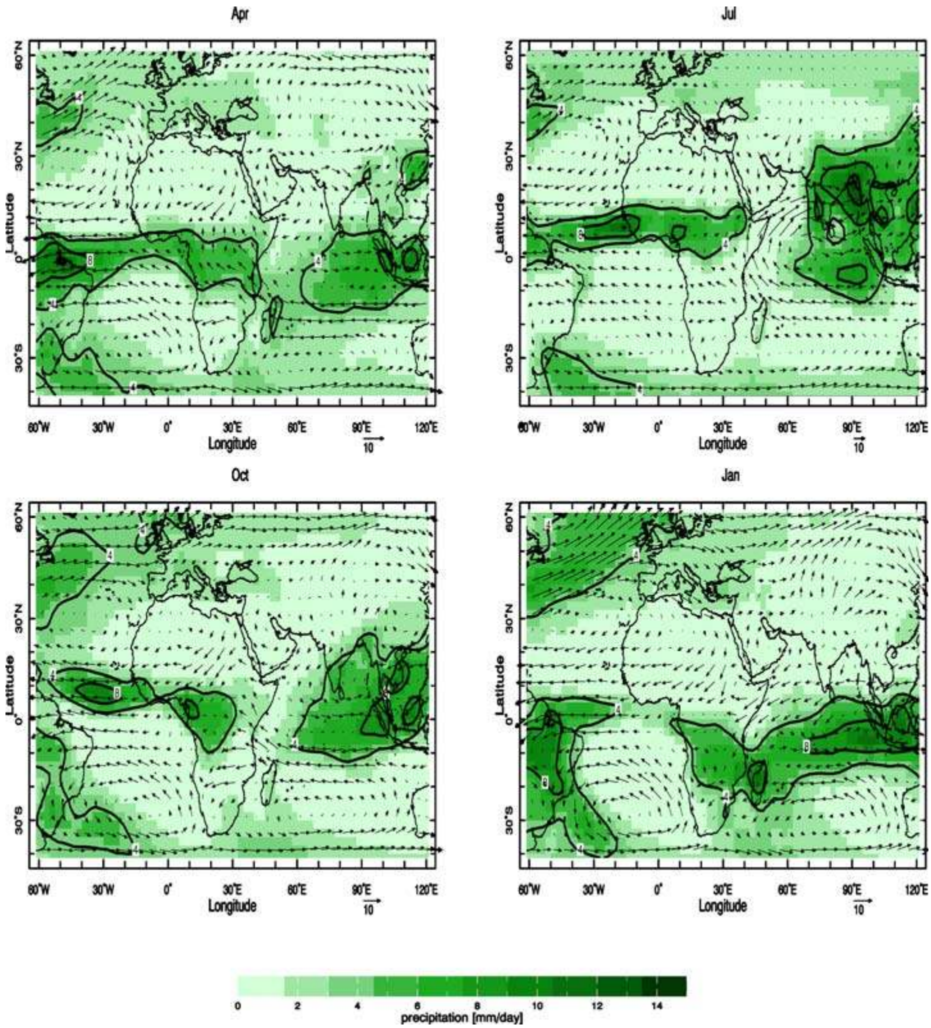

Fig. 1 Climatologies of precipitation and near-surface winds for a domain centered on tropical Africa. Precipitation was produced by the Global Precipitation Climatology Project (Huffman et al. 1997) combining satellite estimates and rain-gauge measurements (contours are drawn every 4 mm/day). Winds at the 850 hPa pressure level—approximately 3,000 m above sea level—are from the NCEP–NCAR Reanalysis Project (Kalnay et al. 1996). Panels are for April, July, October and January averages over 1979–2004

seasonal climatology of African precipitation is shown in Fig. 1; heavy rain occurs in a zonally symmetric,[2] meridionally confined band, or rainbelt, which straddles the equator near the equinoxes, in April and October, and moves off into the

[2]Zonal refers to the longitudinal, or east-west direction, while meridional refers to the latitudinal, or north-south direction. Likewise, zonally or meridionally symmetric refers to a feature that characterizes all longitudes or latitudes in a given domain.

summer hemisphere otherwise, to northern Africa in July and to southern Africa in January. Deep convection and precipitation are spatially confined. In regions of strong precipitation near-surface convergence brings in moisture from surrounding regions, allowing the rainfall to greatly exceed evaporation from the local surface. For this reason, some regions of deep convection are also referred to as Inter-Tropical Convergence Zones, or ITCZs.

Continental surfaces respond immediately to the seasonal cycle of insolation, because of the negligible thermal capacity of land. Hence land surface temperatures peak right after the passing of the sun overhead, or after the local summer solstice at tropical latitudes. Ocean surfaces respond with a delay, due to the large thermal capacity of the layer of water through which this heat is immediately mixed, hence they are typically warmest close to the fall equinox. As the rainbelt progresses poleward in its annual migration following the sun, near-surface winds supply moisture to it, mostly from its equatorward side. As the rainbelt retreats, dry winds, e.g. the Harmattan winds of the Sahara/Sahel, establish themselves on its poleward side. A large-scale circulation connects near-surface convergence and ascent localized in the rainbelt, to upper-level divergence and generalized descent elsewhere. This meridional migration of the rainbelt produces a single-peaked rainy season at the poleward edges of the tropical region. Due to the additional, non-negligible influence of the ocean, peak rainfall occurs with a delay compared to peak temperature (Biasutti et al. 2003), i.e. sometime between summer solstice and fall equinox—in August in the northern hemisphere, and in February in the southern hemisphere. At latitudes closer to the equator, the rainy season is double-peaked, with a dry season in winter, maxima in spring and fall, and a mid-summer break in between. Comparing the July and January panels in Fig. 1, one also notes the seasonal reversal of the prevalent near-surface wind direction typical of monsoon climates, i.e. those climates characterized by a single rainy season and a prolonged dry season.

The role of insolation as a fundamental driver of African climate has been highlighted by paleo-climate studies. The paleo-climatic record unquestionably shows a “green Sahara” in past periods, so much wetter than today that it could sustain a savanna ecosystem in the Tassili-n’Ajjer mountain range of Algeria, located at 25°N, as recently as 6,000 years ago (see Brooks 2004 for a review, and references therein; Petit-Maire 2002). This is explained by an increase in summer insolation due to periodic variations in the Earth’s orbit. Due to the increase in summer insolation, monsoonal rains are driven northwards further into subtropical latitudes. These paleoclimates provide valuable tests of African climate’s sensitivity to changes in insolation. Their simulation by climate models has shown the importance of feedbacks in the oceans and in terrestrial vegetation (e.g. Braconnot et al. 1999).

The potential for changes in insolation to alter African climate also inspired early attempts to explain African rainfall variability in terms of “desertification” (Charney 1975). A key element in this picture is that human activity reduces vegetation cover, which increases the albedo (i.e. reflectivity) of the surface, resulting in less absorbed solar radiation. In analogy with theories of the green Sahara, this reduction in absorbed energy in turn results in a reduction in precipitation. This paradigm for African climate change emerged shortly after inception of the Sahel drought, in the early 1970s, and has remained in the popular imagination since. In the climate science community the desertification paradigm has in fact been supplanted by an alternative that is in many respects its exact opposite—that African rainfall variability is primarily driven by changes in remote sea surface temperatures, rather than in local properties of the land surface itself. Changes in terrestrial vegetation act mainly as amplifiers of the variability forced by the oceans.

3 African climate of the twentieth century

In this section we first describe temporal (Section 3.1) and spatio-temporal (Section 3.2) patterns of observed annual mean African rainfall variability, and sketch their relation to relevant literature on regional climate. We then compare these patterns with those simulated by many coupled ocean-atmosphere models forced by the twentieth century trajectory of anthropogenic aerosol and greenhouse gas emissions (Section 3.3).

We base our observational analyses[3] on the gridded precipitation dataset produced by the Climatic Research Unit of the University of East Anglia (CRU; Hulme 1992). Our baseline analyses cover the period 1930–1995, though in Section 3.1 we extend temporal coverage to the present, using Global Precipitation Climatology Project (GPCP; Huffman et al. 1997) data, a blended satellite-rain gauge product with global, i.e. land and ocean, coverage, available starting in 1979. The period from the 1930s to the mid-1990s affords an appropriate observational coverage over Africa. Coverage as measured by the number of stations recording monthly in the archives of the Global Historical Climate Network (GHCN; Vose et al. 1992), stored at the U.S. National Oceanic and Atmospheric Administration's National Climate Data Center, is best in mid-century, and declines towards the beginning and end (not shown).

3.1 Regional variability in time (1930-2005)

Based on considerations of seasonality—which distinguishes monsoonal from equatorial climates (see previous section)—and of regional political entities, we define three sub-regions; western (0°to 20°N, 20°W to 20°E), eastern equatorial (10°S to 10°N, 20°E to 50°E), and southern Africa (25°S to 10°S, 20°E to 40°E). These regions are broadly consistent with those chosen by Hulme et al. (2001), who also presented the "state-of-the-art" in understanding of African climate as background to consideration of climate change scenarios. The history of annual-mean (July to June) rainfall anomalies averaged over these regions is displayed in Fig. 2.

Comparison of the three panels in Fig. 2 highlights the qualitative difference between the West African time series on one side, and its eastern equatorial and southern African counterparts on the other. West African rainfall is characterized by a high degree of persistence, of anomalously wet (e.g. in the 1930s, 1950s and 1960s) and dry (e.g. in the 1970s and 1980s) years—arguably, the shift in Sahel rainfall is unparalleled globally, in magnitude, spatial extent and duration (Lamb 1982; Katz and Glantz 1986; Hulme 1996; Nicholson et al. 1998; Trenberth et al. 2007). In eastern

[3]All observational datasets mentioned in this paragraph—CRU, GHCN and GPCP by their acronyms—are available via the International Research Institute for Climate and Society (IRI)'s Data Library http://iridl.ldeo.columbia.edu/index.html.

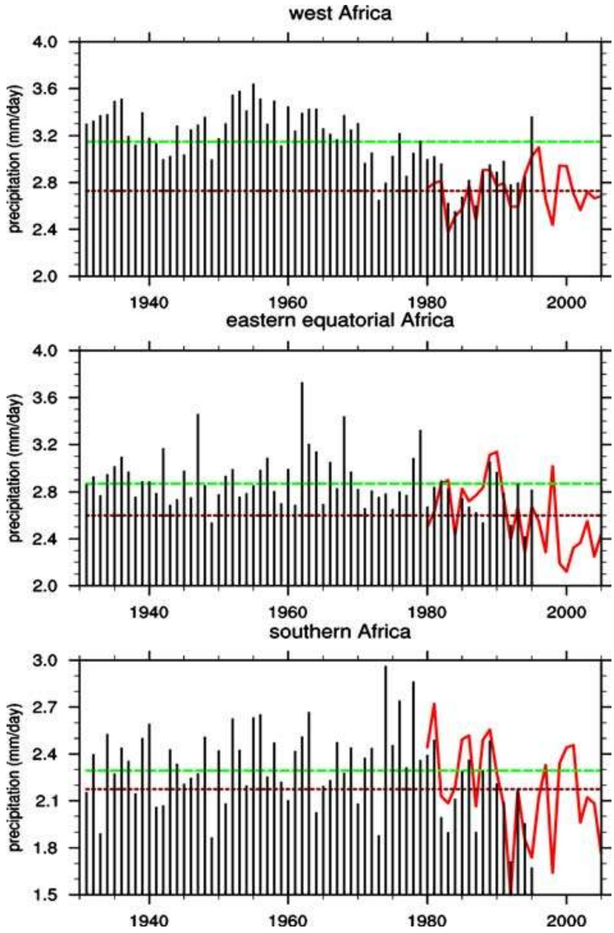

Fig. 2 Regional averages of annual mean (July–June) precipitation over 1930–2005; the three regions are western (0°N to 20°N, 20°W to 20°E), eastern equatorial (10°S to 10°N, 20°E to 50°E), and southern Africa (25°S to 10°S, 20°E to 40°E). CRU data is depicted in *black bars*, GPCP in the *solid red line*. The CRU climatology over 1930–1995 is depicted in the *green horizontal line*, that of GPCP, over 1979-2005, is depicted in the *brown horizontal line*

equatorial and southern Africa interannual variability is more conspicuous. As will become apparent in the next subsection, this has to do with the time scales favored by the oceanic forcings effecting these changes.

A "recovery of the rains" in the West African Sahel is debated in recent literature (e.g. L'Hôte et al. 2002; Nicholson 2005), in relation to a "greening trend" apparent in remotely sensed vegetation since the early 1980s (Herrmann et al. 2005). Simultaneously, eastern equatorial and southern Africa seem to have suffered from more frequent anomalously dry events. The trend towards dry conditions is made even more apparent with the GPCP extension of the time series—the continuous red line in Fig. 2—particularly for eastern equatorial Africa.

3.2 Spatio-temporal patterns of African climate variability

To identify patterns of African climate variability at the continental scale, and to connect these patterns to common, global-scale forcings, we use Principal Component Analysis (PCA; von Storch and Zwiers 1999), a statistical technique widely used in the Earth and social sciences, and linear regression. In our analysis, PCA objectively identifies spatial patterns that express co-variability in time across all tropical African grid points in the CRU dataset. By maximizing the variance captured in the first such pattern, and in subsequent orthogonal, or independent, patterns, PCA identifies broad spatial features, allowing one to summarize a large fraction of the information with a small number of time series. We apply PCA to annual-mean (July-June) precipitation anomalies, with respect to the long-term mean, in the domain between 35°S and 30°N, 20°W and 60°E.[4] We apply linear regression to relate the temporal variability associated with the spatial patterns identified by PCA to other aspects of the global ocean-atmosphere system.

Statistical analysis of African rainfall variability has a long history. In one of the first continental-scale studies Nicholson (1986) identified two recurrent spatial patterns over Africa. One was characterized by continent-wide anomalies of the same sign, the other by anomalies of opposite sign at equatorial latitudes and at the poleward margins of monsoon regions. Nicholson's findings are echoed in our analysis (Fig. 3). Three patterns stand out; one exemplifies a continent-wide drying most significant at the poleward edges of the monsoons, and the other two the influence of ENSO,[5] dominant on eastern equatorial and southern Africa. The three patterns combined capture 37% of the total variance in precipitation over the tropical African domain considered. Our discussion in the following subsections will focus on these patterns. That they are actually representative of regional anomalies is demonstrated by the significant correlation of their associated time series (Figs. 3d, e, f) with the regional averages described in the previous subsection (over 1930–1995); the West African average is unequivocally correlated with the "continental drying" pattern (correlation is 0.97), while eastern equatorial and southern African averages are correlated with both "drying" and "ENSO" patterns (with correlations of the order of 0.5).

3.2.1 A continental precipitation pattern and its relation to global ocean temperature

The leading pattern of annual-mean variability in precipitation over 1930–1995 is displayed in Fig. 3, left column. The spatial pattern (Fig. 3a) is continental in scale. It takes on stronger values in the northern hemisphere, including in the Sahel and along coastal Gulf of Guinea, and weaker, but significant values at tropical latitudes over southern Africa. In fact, this pattern resembles that of the linear trend computed over the same period (not shown)—an overall drying trend. However, the linear trend does not capture all the important features of the time series associated with the pattern shown in Fig. 3. Fluctuations with a multidecadal time scale superimposed on year-to-year variations are evident in Fig. 3d: the 1950s and 1960s were clearly wetter than the long-term average, but so were the 1930s. The wetter-than-average central decades of the twentieth century certainly contribute to making the contrast with

[4]This analysis was repeated on GHCN station data on a shorter period (1950–1995; not shown), yielding consistent results.

[5]A coupled ocean-atmosphere phenomenon which originates in the tropical Pacific, and has world-wide impacts (see Wallace et al. 1998, and accompanying articles in the same issue of *J. Geophys. Res.*). ENSO-related variability can be identified by its preferred interannual time scale, given that an ENSO event usually recurs every 2 to 7 years.

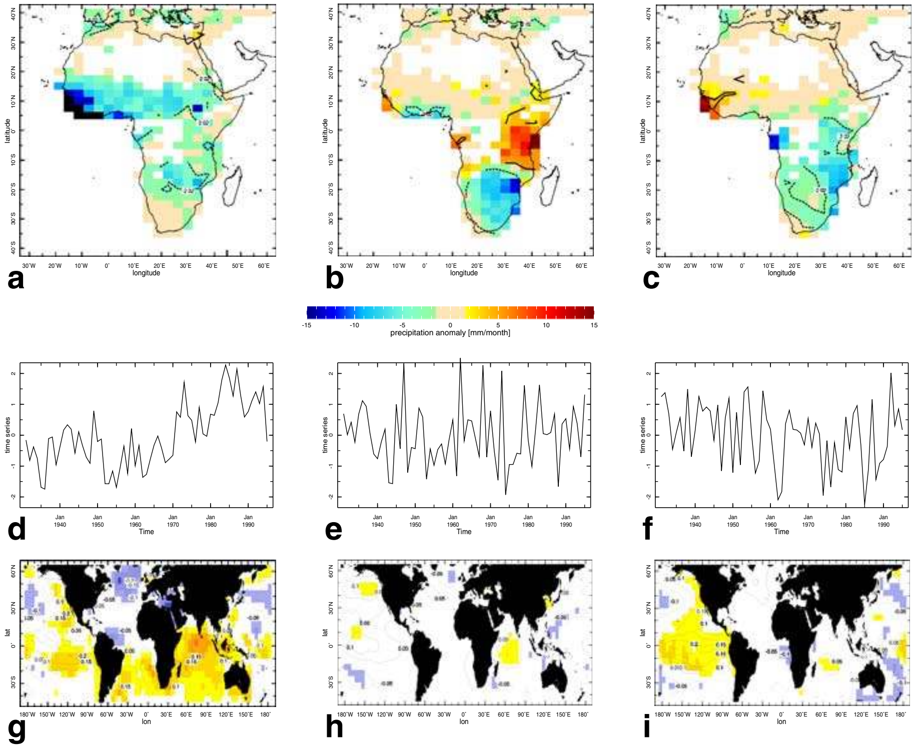

Fig. 3 The three leading patterns of a principal component analysis performed on annual mean (July–June) precipitation over Africa during 1930–1995. **a**, **b**, **c** The spatial patterns, obtained by linear regression of the time series in (**d**), (**e**), and (**f**) onto precipitation—anomalies are in color, in millimeters/month, while the contours delimit regions where regression is statistically significant at the 5% level. **d**, **e**, **f** The associated time series (in units of standard deviation). Note that the upward trend in (**d**) indicates increasingly stronger values of the negative anomalies in the pattern in (**a**), i.e. a drying trend. **g**, **h**, **i** Regression patterns of the time series in (**d**), (**e**), and (**f**) with sea surface temperature (Kaplan et al. 1998). Anomalies are in contour, spaced every 0.05°C, and dashed in the case of negative values; the presence of color represents their statistical significance at a level of 5% or higher

the drier decades that followed, in the 1970s and 1980s, even starker. This pattern is statistically related to global SSTs, with the sign such that drying over Africa is associated with warmer tropical Pacific, Indian and South Atlantic Oceans, and a cooler North Atlantic basin (Fig. 3g; Folland et al. 1986; Giannini et al. 2003).

In the northern hemisphere this drying pattern combines features expressed seasonally in spring and summer respectively along coastal Gulf of Guinea and across the Sahel (not shown; Wagner and DaSilva 1994; Nicholson 1981; Lamb and Peppler 1992; Janicot et al. 1996). Possibly because it is based on annual mean rainfall, our analysis does not isolate the well studied "rainfall dipole pattern" between Sahel and Gulf of Guinea (Fontaine and Janicot 1996; Nicholson et al. 2000; Vizy and Cook 2001, 2002; Cook and Vizy 2006; Reason and Rouault 2006). Previous studies have proposed two explanations for the influence of oceanic temperatures on West African rainfall. The first one is a *moisture supply* argument, articulated by Lamb

(1978) and Folland et al. (1986), and more recently by Rotstayn and Lohmann (2002) and Hoerling et al. (2006). It emphasizes the role of the north/south SST gradient, which is particularly evident in the Atlantic basin (Fig. 3g). This explanation relates the recent, persistent drying of the Sahel to an equatorward shift of the mean location of the Atlantic ITCZ, connecting reduced oceanic precipitation with a weaker West African monsoonal flow. A complementary explanation, consistent with the modeling results of Giannini et al. (2003), Bader and Latif (2003), Lu and Delworth (2005) and Herceg et al. (2007), and with the idealized study of Chiang and Sobel (2002), is a *stabilization* argument in which warmer Indo-Pacific SST causes a warming of the entire tropical troposphere, stabilizing the tropical troposphere from above in the Atlantic sector and over Africa. The key to this mechanism is that the continental boundary layer is unable to increase its energy content and buoyancy to keep up with the energy content over the surrounding oceans. The issue of whether West African rainfall is affected not only by ocean temperature gradients, but also by spatially uniform increases in oceanic temperatures, is central to projections of climate change in the twenty-first century.

In the southern hemisphere the indication for a recent, persistent drying is significant, but not as strong as in the northern hemisphere. When southern Africa is considered in isolation, a significant drying trend is seen since the 1970s (Nicholson 1993; Mason and Tyson 2000), but not on longer, centennial timescales (Fauchereau et al. 2003). As we will show in the next subsection, ENSO is the dominant influence on the predictable component of interannual rainfall variability in eastern equatorial and southern Africa. Nevertheless, anthropogenic influences cannot be ruled out (e.g. Mason 2001). One possibility is that the drying signal in southeastern Africa is due in part to the warming trend in the Indian Ocean (Hoerling et al. 2006), which is understood to enhance the drying impact of warm ENSO events on this region (Richard et al. 2000). Another possibility is that global warming has caused the recent trend towards more frequent and persistent warm ENSO events (Timmermann et al. 1999; Trenberth and Hoar 1997), thus affecting southern African rainfall indirectly. Despite improvements in the simulation of ENSO dynamics in the current generation of coupled ocean-atmosphere models, however, projections of change in either amplitude or frequency of future ENSO events are not robust (Meehl et al. 2007).

3.2.2 Interannual variability related to ENSO

ENSO is the dominant phenomenon in the two remaining patterns (Fig. 3, center and right columns). Its signature in SST—anomalies of one sign centered on the equator in the central and eastern tropical Pacific basin surrounded by anomalies of the opposite sign at subtropical latitudes, in the characteristic "horseshoe" pattern (Rasmusson and Carpenter 1982)—is clear in Figs. 3h, i. ENSO has a well-known impact on rainfall in eastern equatorial and southern Africa (Ropelewski and Halpert 1987; Ogallo 1989; Nicholson and Kim 1997; Goddard and Graham 1999; Schreck and Semazzi 2004; Cane et al. 1994; Unganai 1996; Mason 2001; Mulenga et al. 2003; Reason and Jagadheesha 2005), as depicted in Figs. 3b, c. However, it is also understood to have a non-negligible effect north of the equator (Seleshi and Demaree 1995; Janicot et al. 1996; Ward 1998; Rowell 2001; Giannini et al. 2003; Korecha and Barston 2007), especially when care is taken to separate the year-to-year from the multi-decadal variability described in the previous subsection.

ENSO impacts Africa directly via an atmospheric teleconnection (Glantz et al. 1991; Wallace et al. 1998) and indirectly, via the response of the Indian and Atlantic basins to it (Klein et al. 1999; Alexander et al. 2002). The direct influence of ENSO over Africa follows the path outlined above for the influence of the Indo-Pacific warming trend under the *stabilization* mechanism. During a warm ENSO, or El Niño event[6] the entire tropical troposphere warms as a response to the warming of the central and eastern equatorial Pacific (Yulaeva and Wallace 1994; Soden 2000; Sobel et al. 2002). Outside the central and eastern equatorial Pacific, stabilization of the atmospheric column initially leads to below-average precipitation in remote regions (Chiang and Sobel 2002), as is observed during the growth phase of warm ENSO events, which coincides with the northern hemisphere monsoon season in South and East Asia, Africa and Central America (Ropelewski and Halpert 1987; Kiladis and Diaz 1989; Lyon 2004).

The third pattern captures this purely atmospheric influence—note how the tropical Pacific SST anomalies in Fig. 3i dwarf any signal in the ocean basins around Africa. The atmospheric signal is indeed a drying in the case of a warm ENSO event, as discussed by Goddard and Graham (1999) in the context of comprehensive climate model simulations, and explained by Chiang and Sobel (2002) and Neelin et al. (2003) in more idealized frameworks. In our analysis this signal is strongest in eastern Africa between the equator and 20°S.

The second pattern represents the "canonical" ENSO influence, one that combines the effects of remote, or tropical Pacific, and local, especially Indian Ocean, surface temperatures (Fig. 3h). It captures the wetter than average conditions over eastern equatorial Africa known to be a maximum during the October-December short rains, which coincide with mature warm ENSO conditions, as well as the drier than average conditions over southern Africa known to be most prominent in the January–March season immediately subsequent to the mature phase (Fig 3b). The Indian Ocean SST anomalies associated with the dipolar rainfall pattern between eastern equatorial and southern Africa are related in part to ENSO.[7] Due to their thermal inertia, the remote tropical oceans warm in response to the ENSO-induced changes in the tropospheric temperature with a lag of a few months (see, e.g. Klein et al. 1999; Chiang and Sobel 2002; Sobel et al. 2002; Su et al. 2005; Chiang and Lintner 2005). In some remote oceanic regions, ENSO-induced changes in surface wind stress drive an additional SST increase via ocean dynamics (e.g. Bracco et al. 2005; Song et al. 2007 in the case of the Indian Ocean), reversing the initial direct atmospheric stabilization and leading to destabilization of the tropical troposphere from the bottom up. This effect, together with an increased moisture transport associated with enhanced evaporation over the warmer than average Indian Ocean, can lead to positive precipitation anomalies for eastern equatorial Africa (Goddard and Graham 1999). Hence in this region a warm ENSO can induce below-average precipitation when the atmospheric bridge dominates, as depicted in the third pattern,

[6] To the extent that the system is linear, all features described for a warm ENSO, or El Niño event, are reversed in sign for a cold ENSO, or La Niña event.

[7] The controversy relating to the existence of an Indian Ocean dipole mode in SST independent of ENSO (Saji et al. 1999; Webster et al. 1999), captured in the references to Hastenrath (2002), is beyond the scope of this study.

or it can result in above-average precipitation, when the dynamically-induced Indian Ocean response overwhelms the remote atmospheric effect, as in the second pattern. Song et al. (2007) have recently argued that the timing of ENSO with respect to the seasonal cycle of upwelling in the eastern Indian Ocean helps determine whether the dynamic Indian Ocean component visible in Fig. 3b becomes dominant.

To summarize, when observations of precipitation over Africa are analyzed with a view to their global linkages, two continental-scale patterns, related to variability in the oceans, appear to dominate African climate variability: (1) a continental-scale drying pattern related to enhanced warming of the southern compared to the northern tropics and to a warming of the tropical oceans, (2) the impact of ENSO on the tropical atmosphere and oceans around Africa.

While beyond the scope of this paper, it should be noted that seasonal climate prediction hinges on the exploitation of these relationships: the slow evolution of SST anomalies provides the memory that makes the forecast of rainfall anomalies possible from weeks to a few months out. The potential for practical usefulness has been widely acknowledged. Washington et al. (2006) argue that the development of strategies to cope with climate variability at the interannual time scale should be a primary goal for African climate science. Building adaptive capacity at the faster climate variability time scales, on one hand, can provide a roadmap to adaptation to aspects of climate change (such as in the increase of extreme events) and, on the other hand, can foster the necessary links at the climate-society interface. The necessity for the production, dissemination, and use of climate information has led to the organization of regional seasonal climate outlook forums—an effort that has been on-going for 10 years now in western, eastern equatorial and southern Africa, uniting climate forecasters, stakeholders and media in one venue. Tarhule and Lamb (2003) and Patt et al. (2007) and references therein provide excellent overviews of lessons learned.

3.3 Late twentieth century climate change in recent simulations

In early 2005, simulations with state-of-the-art climate models from 16 research centers were made available to the scientific community by the World Climate Research Program through the 3rd Coupled Model Intercomparison Project (WCRP/CMIP3), hosted at the Program for Climate Modeling Diagnosis and Intercomparison (PCMDI; http://www-pcmdi.llnl.gov/ipcc/about_ipcc.php) in preparation for the 4th Assessment Report of the Intergovernmental Panel on Climate Change (IPCC 4AR), published in 2007. In this subsection and the next section we discuss our analysis of twentieth and twenty-first century simulations of African climate change with 19 models taken from the WCRP/CMIP3 archive (Biasutti and Giannini 2006). Lau et al. (2006a) and Cook and Vizy (2006) provide other valuable analyses of African rainfall using the same archive of simulations. The consensus view is summarized in Christensen et al. (2007).

In this subsection we discuss the comparison between simulations obtained with the same climate models integrated under two different protocols. In one case—the control, or pre-industrial (PI) simulations—aerosol and greenhouse gas emissions are set to constant values representative of conditions before anthropogenic contributions to these climate forcings became significant. In the other—the forced, or twentieth century (XX) simulations—the historical progression of emissions associated with industrialization, which constitute the anthropogenic forcing,[8] is introduced, and influences the evolution of the climate system from the second half of the 19th century to the present.

While the historical concentrations of anthropogenic greenhouse gases are fairly accurately inferred from fossil fuel extraction records, the historical concentration, type, and spatial distribution of aerosols remain uncertain (Forster et al. 2007). The response of the climate system can vary greatly depending on whether the forcing includes only predominantly reflecting aerosols, e.g. sulfate aerosols, or both reflecting and absorbing aerosols, e.g. black carbon, (e.g. Menon et al. 2002; Lau et al. 2006b). Furthermore, only a few models include parameterizations of the indirect effects of aerosols, i.e. those occurring via the aerosols' interaction with cloud microphysics and precipitation efficiency (Lohmann et al. 2000). However, as disparate as the diagnosis or prognosis of the aerosols' impact may be, when the forced, or twentieth century (XX) simulations, are compared to the control, or pre-industrial (PI) simulations, features that are consistent across models can be identified in the surface temperature and precipitation fields.

Multi-model ensemble averages of annual-mean surface temperature and precipitation are presented in the subsections that follow. Multi-model ensemble averaging (see e.g. Krishnamurti et al. 1999; Barnston et al. 2003; Hagedorn et al. 2005 in the context of seasonal climate prediction) is one way to deal with uncertainties related to parameterizations, i.e. those components of the models that represent processes that cannot be simulated from first principles, whether because of finite computational resources or incomplete knowledge of the governing physics. The hope is that differences among models that are not physical will cancel, while common robust responses will be retained. Figure 4 shows the changes in annual-mean surface temperature and precipitation between the last 25 years of the twentieth century (XX) and a 25-year period in the pre-industrial (PI) simulations, i.e. the XX–PI difference (Biasutti and Giannini 2006). The top panels show the average of all models, the bottom panels a measure of inter-model agreement (see caption for details).

3.3.1 XX–PI changes in surface temperature

Compared to pre-industrial times, late twentieth century annual-mean surface temperature increases less in the mid-latitude extratropics than in the deep tropics and in high latitudes (Fig. 4a), with many models cooling the mid-latitude northern hemisphere oceans. This modeled response is very similar to the linear trend pattern in observed surface temperature computed over 1950–2000 (Hansen et al. 1999; http://data.giss.nasa.gov/gistemp/maps/), in both spatial signature and amplitude. This pattern has been the subject of intensive investigation (see, in past IPCC reports, Santer et. al. 1996 and Mitchell et al. 2001, and, more recently, Stott et al. 2006) and we do not discuss it further here except to note that in model simulations reflecting sulfate aerosols contribute to minimizing the warming in the northern hemisphere.

[8]Certain models also include natural variability associated with changes in insolation, and volcanic aerosols, as well as land use/land cover change, in their twentieth century simulations.

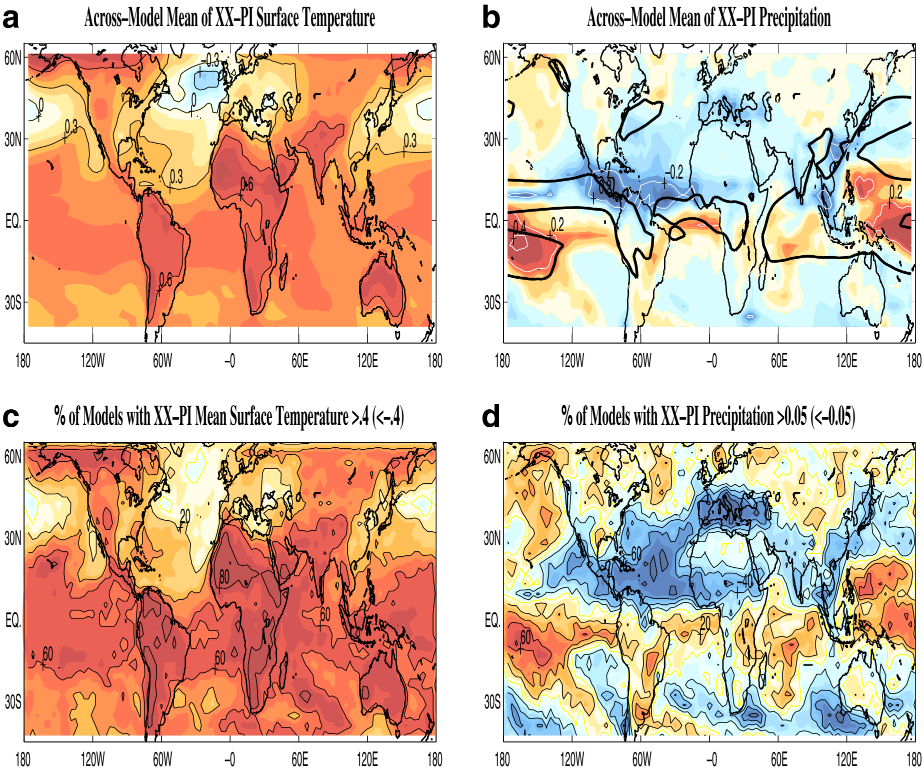

Fig. 4 The difference (XX–PI) between late twentieth century (1975–1999) climate in the XX simulations and pre-industrial climate, as estimated from a 25-year mean from the PI simulations. The left panels refer to annual mean surface temperature, the right panels to annual mean precipitation. *Warm colors* indicate positive and *cool colors* negative anomalies. The *top panels* are the mean across all 19 models (the shading interval is 0.1°C for temperature and 0.03 mm day^{-1} for precipitation), the *solid black line in the right panel* is the 4 mm day^{-1} contour in the XX simulations and indicates the location of mean maximum rainfall . The *bottom panels* are a measure of inter-model agreement: a value of 1 (-1) was assigned at every gridpoint and for every model if the XX–PI difference was positive (negative) and larger than a threshold (0.4° C or 0.05 mm day^{-1}) value; the *panels* show the sum over all models as percentages (shading interval is every 10%, contour interval is every 20%)

3.3.2 XX–PI changes in precipitation

The XX–PI difference in SST discussed above bears a striking resemblance to that identified by Folland et al. (1986) and Giannini et al. (2003) to have been associated with drought in the Sahel (also see Fig. 3g here). Consistently, the response pattern in the annual mean (and, more prominently, in the June–August average) of precipitation shows a decrease across the tropical Atlantic ITCZ and into the Sahel in the late twentieth century compared to the pre-industrial control. This is echoed in all models, except two which show negligible anomalies (Fig. 4d; also see Fig. 2 in Biasutti and Giannini 2006). We conclude that anthropogenic forcings played a non negligible role in the late twentieth century drying of the Sahel (Held et al. 2005; Biasutti and Giannini 2006; Rotstayn and Lohmann 2002), although there is little doubt that a large fraction of the observed variations in Sahel rainfall was generated by natural SST variability. A rough estimate suggests that the anthropogenic contribution is about 30% of the recent long-term drying trend. The current model archive is not suited to separating the effects of aerosols from those of increasing greenhouse gases. The fact that this twentieth century drying trend in the Sahel does not continue into the twenty-first century in a majority of models (see below) nonetheless strongly suggests that models may be more consistently representing a change in the moisture supply mechanism, in this case expressed in the North-South gradient in SST forced by aerosols and attendant displacement of the Atlantic ITCZ, rather than in the stabilization mechanism associated with the more uniform tropical SST warming forced by greenhouse gases (Biasutti and Giannini 2006). However, in one model, that developed at NOAA's Geophysical Fluid Dynamics Laboratory, Held et al. (2005) find a clear signature of drying due to greenhouse gases.

The drying of the Sahel is accompanied, in the annual mean picture, by anomalies in eastern equatorial and southern Africa reminiscent of the response to ENSO, i.e. wet anomalies in eastern equatorial and dry anomalies in southern Africa, but these anomalies are weak and inconsistent across models, even more so during the December-February season (not shown), when the ENSO-related signal should be most prominent. This lack of agreement seems surprising, given that models forced with the historic global SST consistently show a drying trend in southern Africa (Hoerling et al. 2006). It is possible that the observed drying trend was a consequence of the prevalence of warm ENSO events at the end of the twentieth century, and that coupled models do not simulate such prevalence, either because it is not a forced behavior or because the models' simulation of ENSO is not sufficiently accurate.

4 The uncertain future: a global warming scenario simulation

Climate science has made impressive inroads in understanding climate variability and change. A future, further warming of the planet is predicted with certainty. Yet, the regional patterns of climate change, especially potential future shifts in the spatial distribution of tropical rainfall, remain uncertain. In Fig. 5 we compare the climate of the late twentieth century to that of the late twenty-first century as simulated in the A1B climate change scenario by the same 19 models introduced in the previous section—we plot the difference between the last 25 years in the A1B and twentieth century simulations. In the A1B scenario a growing world economy and the adoption of cleaner technologies combine to give growing greenhouse gas concentrations and declining sulfate aerosol loadings.

The global warming signature emerges clearly in the annual-mean temperature difference (Fig. 5a). The models show (1) a more equatorially symmetric warming of the tropical oceans, (2) a more consistent ENSO signal in the twenty-first compared to the twentieth century, with a pronounced, El Niño-like warming of the central and eastern equatorial Pacific,[9] and (3) a more pronounced land-sea temperature

[9] But see Collins et al. (2003), Guilyardi (2006) on the El Niño response to global warming.

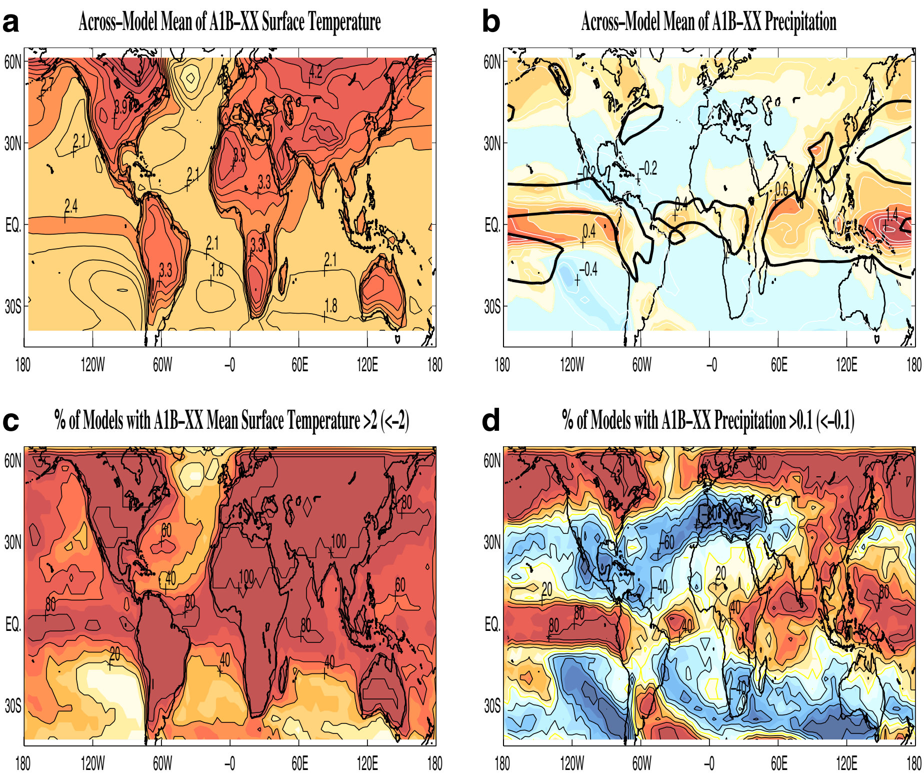

Fig. 5 As for Fig 4, but for the A1B–XX difference. In the *top panels*, the shading (contour) interval is 0.6°C (0.3°C) for temperature and 0.15 mm day^{-1} (0.2 mm day^{-1}) for rainfall. In the *bottom left panel* the threshold is 2°C, while in the *bottom right panel* it is 0.1 mm day^{-1}

contrast, with warming of the continents that surpasses the warming of the tropical oceans. In Africa, the surface warming is largest in desert regions, and relatively more modest in equatorial regions, where the increased radiative forcing is partly used to evaporate soil moisture.

The precipitation signal is less clear. Over Africa, the multi-model mean in precipitation (Fig. 5b) suggests more intense rainfall in equatorial regions and drier conditions elsewhere (Christensen et al. 2007), but this prediction is not unanimous across models (Fig. 5d): it is more robust in eastern equatorial Africa, where a majority of models agree on a trend towards wetter conditions, but less certain in southern Africa, where there is some indication of a drying trend (Hoerling et al. 2006), but much scatter in the model solutions. There is no agreement on whether the Sahel will be drier or wetter in the future (Held et al. 2005; Biasutti and Giannini 2006; Hoerling et al. 2006; Cook and Vizy 2006). Averages for the same three regions defined in Section 3.1—western, eastern equatorial and southern Africa—in the last 50 years of the twentieth century simulations and in the twenty-first century simulations are plotted in Fig. 6, for each model, and for the multi-model mean. The scatter among models, which has not changed significantly since Hulme et al. (2001; see their Figure 13), is the most visible feature.

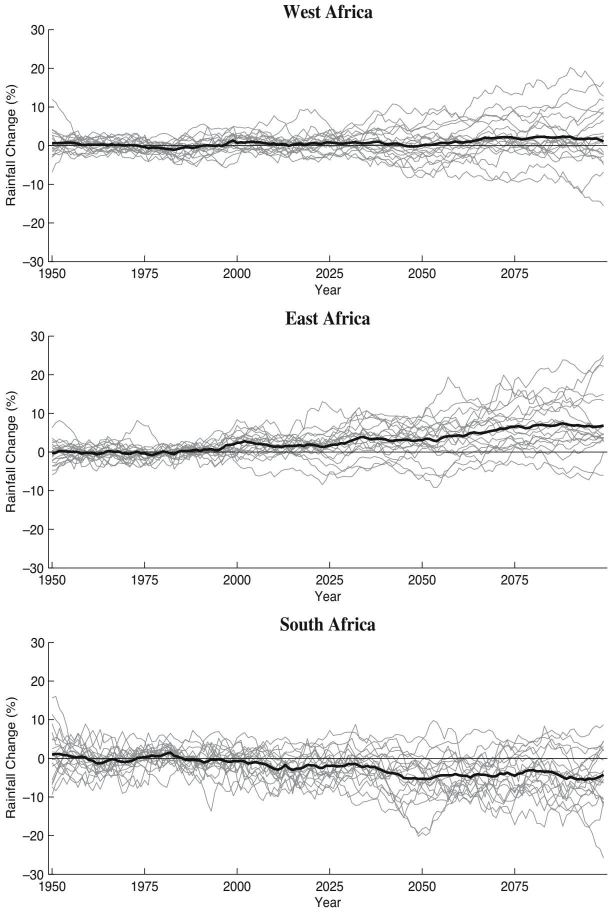

Fig. 6 Regional averages of precipitation in the IPCC 4AR model simulations; twentieth century simulations from 1950 to 2000, and A1B scenario simulations from 2000 to 2100. Regions are defined as in Fig. 2. *Each grey line* represents one model, and the *thicker black line* is the multi-model mean

These results are based on a uniform weighting of all models in the WCRP/CMIP3 archive. But some of the models clearly have better simulations of African rainfall than others (Cook and Vizy 2006; Kamga et al. 2000). The question of how best to weigh the different projections in such an ensemble of models remains open. There are many possible metrics, some focusing on regional climate simulations and others evaluating the models more globally, some on the behavior of the atmospheric model when run over observed SSTs, and others on the behavior of the coupled system. What is lacking at present is an accepted strategy for determining which metrics are most relevant to evaluate a model's projection of African rainfall.

Partly based on inspection of this model archive we can generate hypotheses that may help clarify why models differ in their twenty-first century responses (also see Biasutti et al. 2008, recently revised for *Journal of Climate*). A uniform warming is confidently expected to lead to an increase in specific humidity (Allen and Ingram 2002; Held and Soden 2006). The hypothesis that regional precipitation changes are determined by this humidity change, rather than by changes to the circulation (e.g. Trenberth et al. 2003), leads to a prediction of an increase in rainfall in currently rain-rich regions, and a decrease in currently semi-arid regions, a prediction that is borne out in climate simulations in a zonally averaged sense, but not throughout Africa (although it may help in interpreting the projected increase in rainfall in eastern equatorial Africa). A related response, in which the rainiest regions become rainier, and drying occurs at the margins of those regions, is predicted by a climate model of intermediate complexity, in association with circulation as well as humidity changes (Neelin et al. 2003). (3) El-Niño-like climate change could lead to teleconnections similar to those seen at interannual time scales, which would tend to dry much of the African continent. (4) An enhanced land-sea temperature contrast, whereby oceans warm less than continents due to land surface or cloud feedbacks, could lead to an intensification of monsoonal circulations, hence an increase in moisture supply and thus rainfall in the Sahel (Haarsma et al. 2005). The current challenge is to determine the relative importance of these various mechanisms over Africa, and to design metrics that tell us whether the model simulations are adequate to provide a meaningful prediction of future change in African rainfall.

5 Conclusions: a climate perspective on African environmental change

African environmental change is often interpreted as a regional phenomenon with local anthropogenic causes and potentially global effects. On the one hand, the local causes are commonly ascribed to the mismanagement of limited natural resources under the ever-increasing pressure of a growing, vulnerable population, while external causes of degradation such as remotely-forced drought are underplayed. A perverse feedback loop in which the poorest of the poor deprive themselves by impoverishing the environment they depend on for their livelihoods is often simplistically invoked, despite much literature that expands on the complexity of the environmental degradation-poverty nexus (e.g. Broad 1994; Duraiappah 1998). On the other hand, the global threat posed by African environmental change is often exaggerated. The notion that human-induced change in land use or land cover plays a significant role in altering African rainfall, and from there the global atmospheric circulation, has not been proven (see Nicholson 2000, and references therein for a review). Rather, the evidence summarized in the Introduction, based on remote sensing of vegetation cover and on climate modeling, falsifies the claim that these local processes are quantitatively important compared to the imprint of large-scale patterns of climate variability and change on regional scales.

In this paper we explored a climate perspective on African environmental change by focusing on how global climate can cause large-scale changes in rainfall patterns over Africa. We identified two mechanisms relevant to changes in African rainfall of the recent past and future—a moisture supply mechanism, and a stabilization mechanism—and indicated that competition between them in a warming world may explain the uncertainties in projections of African rainfall into the next century.

Our analysis of annual-mean, continental-scale variability of African precipitation, as observed over the better part of the twentieth century and modeled in the WCRP/CMIP3 simulations, supports the link between a trend towards drier conditions for the monsoon regions of Africa over the second half of the twentieth century and a pattern of tropical ocean warming (see Sections 3.1 and 3.2). The persistence of drying in the Sahel is confirmed locally (L'Hôte et al. 2002; West et al. 2007). We highlight the skill of the WCRP/CMIP3 models in reproducing certain large-scale features (see Section 3.3), and discuss evidence in support of partial attribution of drought in the Sahel to anthropogenic forcings, specifically to the combined effects of aerosols and greenhouse gases (Held et al. 2005; Biasutti and Giannini 2006; Rotstayn and Lohmann 2002). Similarly, we relate a drying trend in tropical southern Africa, consistently found in very recent observations and in projections, to the warming of the Indian Ocean, while agreeing that its potential anthropogenic origin may have been confounded by the more frequent occurrence of warm ENSO events since the mid-1970s compared to the mid-twentieth century.

However, the broad agreement across models on the relationship between global tropical sea surface temperature and African rainfall does not continue to hold in model projections of West African climate (Hoerling et al. 2006; Christensen 2007). The very recent multi-year dry spell observed in eastern equatorial Africa is at odds with projections of wetter equatorial climates. These discrepancies obviously point to work that needs to be done to fully understand the complex interplay between oceanic and continental, local and remote influences of variability and change.

In conclusion, recent advances in climate research confirm the need for a global climate system perspective to interpret the causes of change in African rainfall and their consequences on the environment—a shift in focus from regional land use and land cover and the actions of Africans, to the global distribution of oceanic temperatures, both naturally varying and as influenced by anthropogenic emissions from the industrial north. It is hoped that the global community's "mainstreaming of climate in development", i.e. the inclusion of a global climate change perspective in discussions of African development, will result in renewed consideration of issues of environmental justice and the morality of mitigation vs. adaptation strategies.

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