Biosphere–atmosphere interactions over West Africa. II: Multiple climate equilibria

By GUILING WANG* and ELFATIH A. B. ELTAHIR
Massachusetts Institute of Technology, USA

SUMMARY

This paper presents both theoretical and numerical analyses on the multiple-equilibrium nature of the regional climate system in West Africa. Based on simple analyses on how the coupled biosphere–atmosphere system responds to vegetation perturbations within the scope of a dynamic ecosystem, we propose that the regional climate system may have multiple equilibrium states coexisting under the same precessional forcing. Using a synchronously coupled biosphere–atmosphere model which includes explicit representation of ecosystem dynamics, we show that the equilibrium state of the model is sensitive to initial vegetation distribution. This modelling result supports the existence of multiple climate equilibria. Using the same model, further experiments are carried out to investigate how the coupled system responds to non-permanent vegetation perturbations. Our results demonstrate how transitions between different climate equilibria can take place when governed by the two-way biosphere–atmosphere feedback. These findings advance our understanding regarding the mechanisms of climate variability over West Africa.

KEYWORDS: Biosphere–atmosphere interactions Climate transition Drought Ecosystem dynamics Multiple climate equilibria West Africa

1. INTRODUCTION

Since its onset in the late 1960s, the severe persistent drought in West Africa has drawn considerable research attention. Motivated by the ideas introduced in Charney's (1975) bio-geophysical feedback hypothesis, numerous studies have been devoted to the climatic impact of land-cover changes, and related the current Sahel drought to the anthropogenic vegetation degradation over West Africa (e.g. Charney *et al.* 1977; Sud and Molod 1988; Xue and Shukla 1993; Zhang and Henderson-Sellers 1996; Zheng and Eltahir 1997, 1998). Most of these studies suggested that the positive feedback triggered by a change in land cover is a significant mechanism for drought persistence. The climate change due to a vegetation perturbation might be sufficient to prevent the vegetation from growing back, thus making the drought self-perpetuating. However, due to the lack of representation of ecosystem dynamics in previous climate models, studies on this land-surface feedback have been limited to how the atmospheric climate responds to prescribed vegetation changes. The related topic of how the perturbed vegetation responds to the induced climate change has not been considered in a fully interactive way. The cycle associated with Charney's feedback hypothesis has not been closed. It still remains to be demonstrated whether the land-cover change does induce such a positive feedback as suggested, and if so, under what circumstance.

The man-made desertification and deforestation over West Africa might have contributed to the current drought, but simply attributing the current drought to human activities fails to look at the present climate in the context of the long-term climate variability. Although the current drought is unprecedented in both severity and persistence, West Africa experienced similar droughts in the past several centuries, long before the human activity started significantly to impact the ecosystem in that region (Maley 1973, 1981; Nicholson 1981; Farmer and Wigley 1985). Historical records of the fluctuation in the level of Lake Chad (Fig. 1) show that wet and dry spells alternate in West Africa. Even during the current century, the climate in most parts of West Africa is

* Corresponding author: Ralph M. Parsons Laboratory, 48-208, Massachusetts Institute of Technology, Cambridge, MA 02139, USA. e-mail: glwang@mit.edu
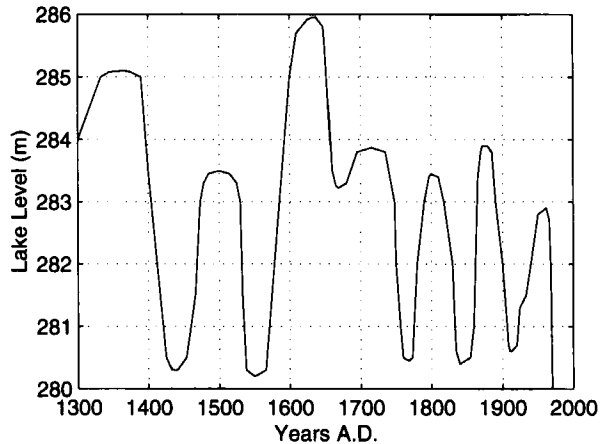
Figure 1. Historical level of Lake Chad (13.5°N), reconstructed based on Farmer and Wigley (1985), Maley (1973), and Nicholson (1981).

dominated by low-frequency variability (Rowell *et al.* 1995). The historical occurrence of dry and wet spells implies that the observed low-frequency climate variability in West Africa may be a natural phenomenon independent of human activities. In this paper we argue that the climate system has multiple equilibrium states coexisting under the same precessional forcing, and interpret the low-frequency climate fluctuations in West Africa as a collective reflection of climate transitions between different equilibria. Here, the multiple equilibria are of deterministic nature, therefore are different from the 'preferred modes' found by statistical–dynamical studies (Nicolis 1982; Demaree and Nicolis 1990; Rodriguez-Iturbe *et al.* 1991; Entekhabi *et al.* 1992; Wang *et al.* 1997) which viewed climate as a stochastic process.

The possibility of multiple equilibria has been suggested by several previous studies (e.g. Nicholson 1989; Entekhabi *et al.* 1992), but lacks clear demonstration prior to this study. A series of unique studies on this topic were carried out by Claussen (1997, 1998), who used the asynchronously coupled ECHAM-BIOME model to investigate the sensitivity of the model's equilibrium to its initial vegetation conditions. The ECHAM-BIOME model found two equilibrium solutions in North Africa and Central East Asia. However, BIOME is an equilibrium vegetation model which cannot simulate ecosystem dynamics, and the coupling between ECHAM and BIOME is asynchronous. These two limitations make it difficult to address the issue of multiple equilibria which is closely related to the synchronous biosphere–atmosphere coupling as well as ecosystem dynamics. In addition, using an equilibrium vegetation model, it would be impossible to address whether and how the climate can actually make a transition between different equilibria. Nevertheless, Claussen's (1997, 1998) were the first modelling studies on the concept of multiple equilibria between the biosphere and the atmosphere.

In this paper, the issue of multiple equilibria is addressed using a synchronously coupled biosphere–atmosphere model which includes explicit representation of ecosystem dynamics. First, we present a theoretical hypothesis on the existence of multiple climate equilibria based on the role of ecosystem dynamics. Then, using the Zonally symmetric, synchronously coupled Biosphere–Atmosphere Model (ZonalBAM) introduced in the companion paper (Wang and Eltahir 2000), we investigate the model sensitivity to initial vegetation distribution and the response of the climate system to non-permanent vegetation perturbations. Our results clearly demonstrate that the climate system indeed

has multiple equilibrium states, with perturbation-induced transitions between different equilibria. The low-frequency climate fluctuations over West Africa are then discussed based on the multiple-equilibrium nature of the regional climate system.

2. HYPOTHESIS

Vegetation plays a prominent role in the exchange of energy, moisture, momentum and carbon between the land surface and the atmosphere. Removal of vegetation modifies the local energy balance and the local water cycle. According to Eltahir (1996), the deforestation-induced reduction in net radiation causes a decrease in the total heat flux from the surface, which is the source of entropy for the atmospheric boundary layer. As a result, the boundary-layer entropy over the land decreases, which favours a weaker monsoon circulation and less monsoon rainfall over West Africa (Eltahir and Gong 1996); therefore, a drier climate results from vegetation degradation. However, this response does not necessarily point to a positive feedback. Whether a feedback is positive or negative depends not only on the climate response to the vegetation change, but more importantly, on the vegetation response to the induced climate change. Here we define a feedback as negative if it moderates the perturbation, and as positive if it enhances the perturbation.

In moist regions such as the tropical rain forest, it is the energy availability, instead of the water availability, that limits the plant growth. Therefore, a rainfall decrease within a certain range has negligible effect in limiting the plant growth. Even in arid regions such as the Sahel, where plant growth is limited by water availability, a reduction of rainfall caused by desertification still does not necessarily trigger a positive feedback. With the degradation of vegetation, the water demand of the vegetation community would also decrease. Most likely, rainfall following the vegetation degradation may not fulfil the water requirement to maintain the pre-perturbation vegetation. However, it may still be enough for the maintenance and further growth of the post-perturbation vegetation. Under this circumstance, the biosphere–atmosphere feedback will be negative, and will drive the system towards its pre-perturbation state, although sometimes it may not fully recover. Otherwise, if the reduced rainfall cannot support the post-perturbation vegetation, the biosphere–atmosphere feedback will be positive, and the vegetation degradation will be enhanced and therefore self-perpetuating.

Based on the above analysis, when a biosphere–atmosphere system at equilibrium is perturbed, the system may respond in three qualitatively different ways: a negative feedback leading to a full recovery; a positive feedback leading to a perturbation enhancement; a negative feedback leading to a partial recovery. When a positive feedback takes place, or when a negative feedback only leads to a partial recovery, the biosphere–atmosphere system will develop into a different equilibrium state. Therefore, we propose that the biosphere–atmosphere system can have multiple equilibrium states coexisting under the same external forcing (due to the orbital motion of the earth), with the two-way biosphere–atmosphere feedback acting as the mechanism for both the climate persistence at one equilibrium and the climate transition between different equilibria.

A system with multiple equilibria can be illustrated using the movement of a ball on a simple landscape, as shown in Fig. 2. When a perturbation occurs, the response of the system depends on both the perturbation magnitude and the status of the post-perturbation system relative to its neighbouring equilibria. With perturbation of increasing magnitude, the system falls into different response zones as shown in Figs. 2(a)–(c) in sequence, which represent the case of a full recovery, a perturbation enhancement, and a partial recovery, respectively. The climate system can remain around
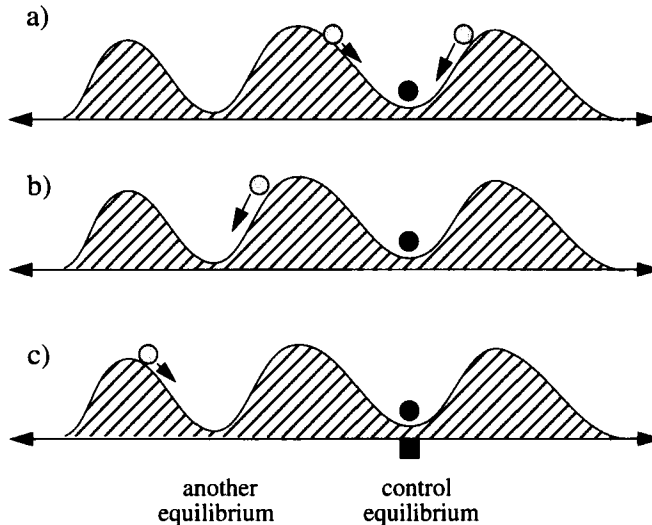
Figure 2. The movement of a ball on a simple landscape, as an analogy of the climate system with multiple equilibria. The filled circle represents an equilibrium climate system, and the shaded circles mark the extent of the perturbations imposed on the system. The three cases represent the system's response to different perturbations: (a) small perturbations—a negative feedback leads to a full recovery, (b) larger perturbations—a positive feedback leads to a new equilibrium, and (c) perturbations even larger—a negative feedback leads to a new equilibrium. As the perturbation magnitude increases further, (b) and (c) will take place alternately.

one equilibrium under the influence of small perturbations (i.e. climate persistence) until a large enough perturbation leads the system towards a different equilibrium (i.e. climate transition). It is worth emphasizing that, when the perturbation is large enough, the transition to a different equilibrium may take place via a negative feedback which leads to a partial recovery.

Our hypothesis not only applies to anthropogenic vegetation perturbations, it also applies to natural variations in the large-scale atmospheric climate forcings. The biosphere–atmosphere system responds to both types of disturbances in a similar way. A large disturbance may develop into a persistent anomaly, while a small disturbance may be wiped out through the two-way feedback between the biosphere and the atmosphere.

In the following we use our modelling results to support the above theoretical hypothesis.

3. MODEL SENSITIVITY TO INITIAL VEGETATION DISTRIBUTION

One necessary condition for a system to have multiple equilibria is its sensitivity to initial conditions. Using ZonalBAM, experiments are carried out to test the sensitivity of the synchronously coupled biosphere–atmosphere system to its initial vegetation distribution. The two extremes of initial vegetation distribution, a uniform forest-covered West Africa and a uniform desert-covered West Africa, are considered.

The land–ocean boundary is set at 6°N. The Atlantic sea surface temperature is fixed at its climatology (1950–1979) (Reynolds and Smith 1994), averaged over 15°W–15°E. Since a zonally symmetric model cannot correctly simulate the atmospheric circulation in mid-latitudes, nor, therefore, the related biosphere–atmosphere feedback, we only apply the vegetation dynamics to the region of 6°N–27°N. Under the extreme initial vegetation distributions we are considering, the simulated climate is expected
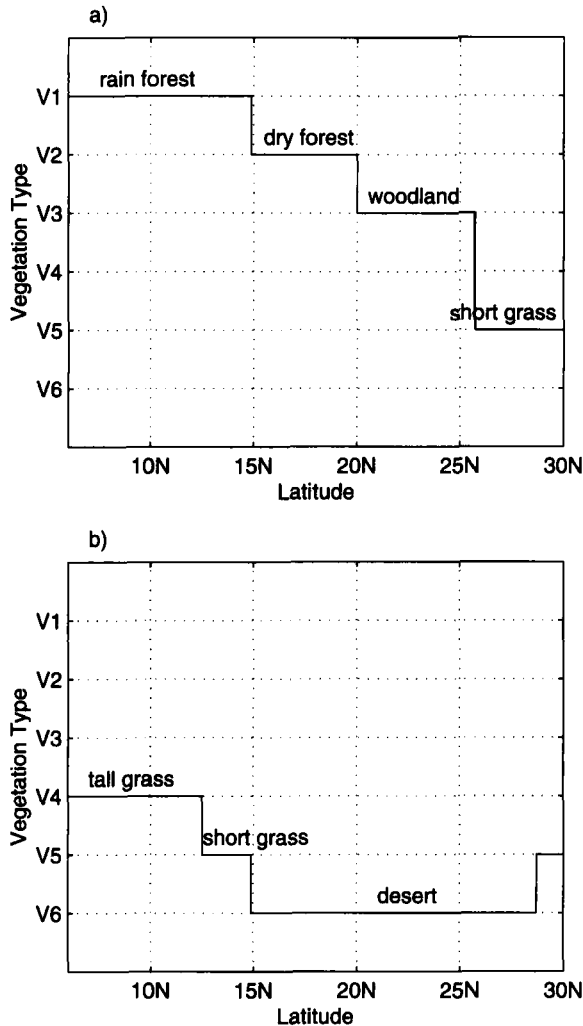
Figure 3.   Vegetation types at the equilibrium state that the coupled biosphere–atmosphere model develops into when it starts with two extreme initial conditions: (a) the model starts from a rain-forest-covered West Africa, and (b) the model starts from a desert-covered West Africa. The vegetation types include V1 (rain forest), V2 (dry forest), V3 (woodland), V4 (tall grass), V5 (short grass), and V6 (desert).

to be very different from the current climate not only within but also beyond the Tropics. Therefore, instead of fixing the land surface fluxes and other properties beyond the Tropics at their climatological values (Wang and Eltahir 2000), here we fix the vegetation north of 27°N at short grass and let the model compute the surface fluxes and temperature. The model results show very little sensitivity to the vegetation condition beyond the Tropics, due to the lack of interactions between the Tropics and mid-latitudes in a zonally symmetric atmospheric model.

Starting with each of the two extreme initial vegetation distributions, the synchronously coupled model is run until it reaches equilibrium. Comparison between these two experiments shows that the modelled biosphere–atmosphere system is indeed sensitive to its initial vegetation distribution. When starting with forest everywhere, the model evolves into an equilibrium (F.I.C.) with trees covering most of West Africa and grasses
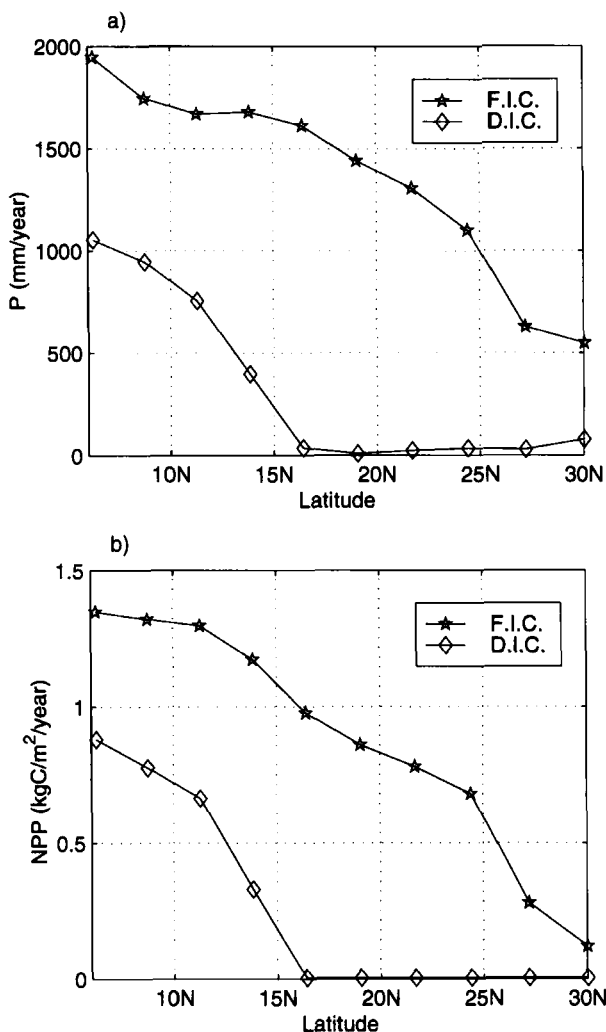
Figure 4. The equilibrium state that the biosphere–atmosphere model develops into when starting with two extreme initial conditions: (a) annual precipitation, and (b) net primary productivity. F.I.C. stands for the case initialized with a rain-forest-covered West Africa, and D.I.C. for the case initialized with a desert-covered West Africa.

covering a narrow band in the north (Fig. 3(a)). When starting with desert everywhere, the vegetation at equilibrium (D.I.C.) varies from tall grass near the coast to short grass and desert northwards (Fig. 3(b)). Correspondingly, equilibrium F.I.C. is significantly wetter and more productive than equilibrium D.I.C., as shown by the distribution of annual rainfall (Fig. 4(a)) and net primary productivity (NPP) (Fig. 4(b)).

Due to the extreme nature of the two initial vegetation distributions, we would expect that the equilibrium F.I.C. and D.I.C. respectively would represent the wettest (the most productive) and driest (the least productive) equilibrium of the natural biosphere–atmosphere system under the current external orbital forcing. Other equilibria of the climate system will most likely fall in between. This statement only holds at the continental scale. As will be demonstrated later, it is possible for this regional climate system to have an equilibrium which is slightly wetter than the F.I.C. equilibrium in a
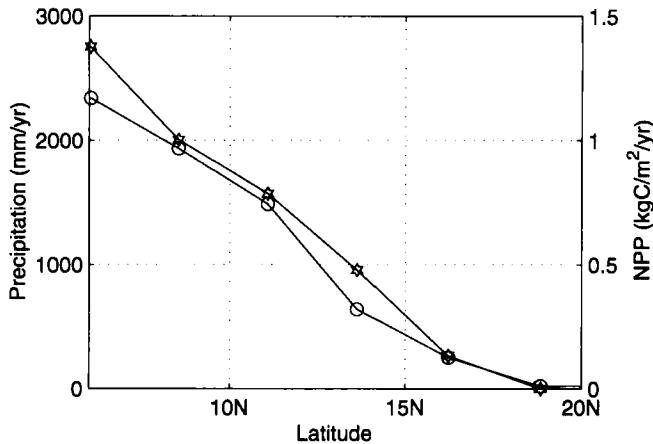
Figure 5.   Precipitation (solid line with open circles) and net primary productivity (solid line with hexagrams) at the control equilibrium state. This equilibrium is derived by starting the model with a vegetation distribution that is close to present day conditions.

small subregion but much drier in the rest of the region. Overall, at the continental scale, the F.I.C. equilibrium is by far the wettest and the most productive.

In principle, the sensitivity of the model equilibrium to initial vegetation distribution already demonstrates the existence of multiple equilibria. However, it is not obvious whether the transition between different equilibria is physically possible. If the regional climate could not evolve from one equilibrium to another under the same external forcing within a reasonable time, the existence of multiple equilibria would be practically irrelevant to the observed climate variability over West Africa. To address the issue of climate transition, experiments have been carried out to investigate the resilience of the biosphere–atmosphere system with respect to different types of perturbations, as presented in the following.

4. RESPONSE OF THE COUPLED SYSTEM TO PERTURBATIONS

In studying the resilience of the biosphere–atmosphere system, we focus on the equilibrium that is close to today's climate condition. Here, the control equilibrium is the equilibrium state derived from 40 years of simulation using ZonalBAM initialized with today's land surface condition, as presented in the companion paper. The annual precipitation and NPP of this control equilibrium, which is close to, but slightly wetter and greener than, today's climate, are presented in Fig. 5.

With respect to the control equilibrium, major land-surface changes of different type and magnitude are added to the system in the 41st year of the simulation. Our study focuses on the impact of non-permanent perturbations—a perturbation takes place during the 41st year, then the landscape is allowed to evolve naturally through biosphere–atmosphere feedback. Three different types of perturbations are considered: deforestation, desertification, and irrigation. Here the deforestation and desertification take place as uniform thinning instead of mosaic exposure of bare soil. Similarly, the effect of irrigation is simulated by a uniform increase of vegetation density.

(a) Deforestation

Deforestation is imposed on the forest and woodland region (6°N–12.5°N) of the control equilibrium. A group of experiments has been performed, with deforestation

ranging from mild selective logging (tree thinning) to forest clearing (the most severe form of deforestation). During the early stage of succession, in all these deforestation experiments, major precipitation reduction takes place. However, the biosphere–atmosphere system can always fully recover from the deforestation perturbation. Here we only present the case of forest clearing as an example. In this experiment, a clear cut removes all the tree biomass over the whole forest and woodland region, leaving behind only the sparse ground vegetation that usually grows during the process of tree clearing.

Figure 6(a) shows the distribution of annual precipitation over West Africa before and after the clearing; Fig. 6(b) and (c) present the growing season leaf area index (LAI) for the upper and lower vegetation canopy, respectively. In the first year following the perturbation, rainfall over the perturbation zone decreases dramatically (Fig. 6(a)). This effect of deforestation on the climate system is not limited to the perturbation zone— the grassland region in the north also experiences a certain degree of rainfall reduction during the year of the perturbation (Fig. 6(a)). Although deforestation reduces the local rainfall by 30–40%, the post-perturbation rainfall is still as high as 1000–1500 mm y^{-1}, which is more than enough for the herbaceous ground vegetation to grow and prevail. As shown in Fig. 6(c), the sparse herbaceous vegetation becomes overwhelmingly dense within one year after the forest clearing. The quick colonization of herbaceous plants following a large-scale clear cut is consistent with ecological observations in the Tropics. "*Succession in the wet tropical lowlands usually starts with rapid soil coverage by a mixture of weedy herbaceous plants and fast growing vines*" (Ewel 1983). These dense herbaceous plants remain as the dominant vegetation for decades until trees eventually take over (Figs. 6(b) and (c)). Interestingly, the grassland region north of the perturbation zone enjoyed a very wet period (Fig. 6(a)) after the establishment of dense herbaceous plants in the deforested region and before the eventual take-over by the upper canopy. Correspondingly, the vegetation growth over the grassland region is enhanced during that period, as shown by the increase of growing season LAI in Fig. 6(c). After the recovery of the upper canopy over the perturbation zone (judged from the upper canopy LAI), rainfall over the entire domain returns to its pre-perturbation level (Fig. 6(a)).

The time it takes for the upper canopy to recover fully ranges from one decade near the coast to four decades at the northern boundary of the perturbation zone. Here the recovery of the upper canopy is measured using upper-canopy LAI. To avoid confusion, it is worth pointing out that although the LAI and NPP reach the pre-perturbation level within years after the selective logging and within decades after a clear cut, it takes centuries for the carbon storage to recover due to the slow accumulation of woody biomass. According to Ewel (1971) and Golley (1975), leaf production is very high during the early stage of forest succession. The LAI for a secondary stand of six years old can be as large as the LAI for a well-stored mature forest (Golley *et al.* 1975). It is the leaf area, not the woody biomass, that really matters for the biospheric production and the biosphere–atmosphere interaction. Therefore, in this study we define the biospheric state based on the LAI instead of the carbon biomass.

Judged from the evolutionary pattern of precipitation and upper-canopy LAI within the perturbation zone (Figs. 6(a) and (b)), it is readily concluded that the deforestation-induced biosphere–atmosphere feedback is negative, and this negative feedback leads to a full recovery. The biosphere–atmosphere system reaches its pre-perturbation equilibrium within four decades following the forest clearing. The recovery time is much shorter following selective logging. The event of rainfall enhancement in the grassland region during the process of forest succession is not observed in the selective-logging experiments, when the deforestation is not severe enough to allow the development of a dense ground vegetation on the forest floor.
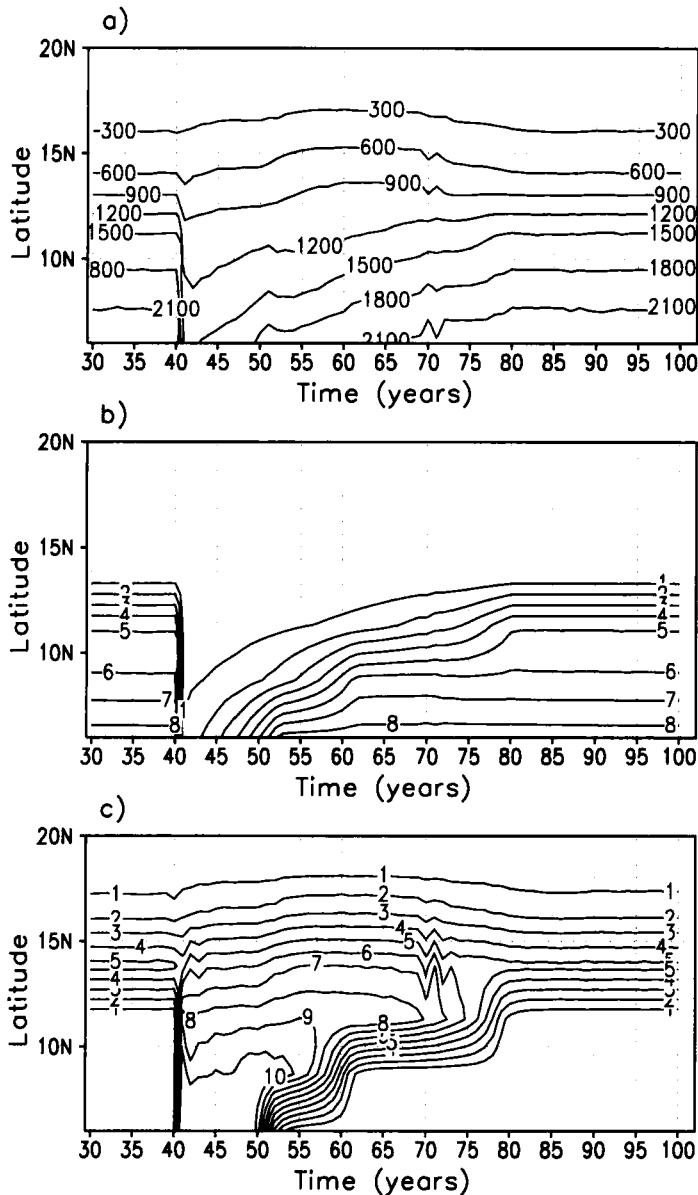
Figure 6.   Evolution of the biosphere–atmosphere system before and after a forest clearance that takes place in the 41st year over the entire forest and woodland region: (a) annual precipitation (mm y^{-1}), (b) growing-season leaf area index (LAI) for the woody plants, and (c) growing-season LAI for the herbaceous plants.

(b)   Desertification and irrigation

The modelled desertification takes place at the grassland region of the control equilibrium, between 12.5°N and 17.5°N. The vegetation degradation is in the form of homogeneous grass thinning, which resembles the effect of grazing. A group of experiments has been carried out, with the severity of desertification increasing gradually. Compared with the system's response to deforestation, its response to desertification is more diverse—desertification of different magnitude can cause qualitatively different

biosphere–atmosphere feedback. Here we present the results from four specific ex-periments Exp1, Exp2, Exp3 and Exp4. These four are chosen because their induced biosphere–atmosphere feedbacks are close to the borders between the responses of dif-ferent types. In Exp1–Exp4, 50%, 60%, 75%, and 80% of the grass biomass are uni-formly removed from the entire perturbation zone, and the induced exposures of bare soil are about 25%, 40%, 60%, and 75%, respectively.

The impact of desertification on the biosphere–atmosphere system is rather local-ized. In spite of the significant reduction in precipitation as well as NPP over the per-turbed region immediately after desertification, the induced precipitation change over the forest and woodland region is relatively small compared to the large rainfall amount before the perturbation. Correspondingly, vegetation over the forest and woodland re-gion shows little detectable response. Therefore, in the following analyses, we focus on the response of the climate system over the grassland region, i.e. the perturbation zone, and take the grid point near 16°N as an example. For all four experiments (and also Exp5 which is described later), the annual precipitation in each year before and after the perturbation is plotted in Fig. 7(a), and the growing-season LAI is plotted in Fig. 7(b). The evolutionary pattern is similar in other parts of the perturbation zone as well as for other properties of the coupled system.

As shown by Figs. 7(a) and (b), after removing 50% of the grass biomass (Exp1), the biosphere–atmosphere interaction triggers a negative feedback, which drives the system back to its pre-perturbation equilibrium within ten years. However, when the fraction of grass removal increases to 60% (Exp2), a positive feedback takes place, which leads to a perturbation enhancement—both the precipitation and LAI further decrease with time after the perturbation. The biosphere–atmosphere system evolves into a different equilibrium state which we define as New1. As the perturbation magnitude increases to 75% grass removal (Exp3), a negative feedback is encountered again, but only leads to a partial recovery—the system rebounds to a certain degree and evolves into the same equilibrium state as in Exp2. As we further increase the grass loss to 80% (Exp4), the biosphere–atmosphere feedback becomes positive again, which drives the system into a more arid equilibrium state, defined as New2. The equilibrium New2 features desert conditions at the grid point near 16°N where grassland exists at the control equilibrium. Induced by different degrees of desertification over the region 12.5°N–17.5°N, the biosphere–atmosphere system is able to evolve from the control equilibrium into two new equilibria.

To simulate the effect of irrigation, several experiments have been performed on the uniform increase of grass density over the grassland region 12.5°N–17.5°N. As an example, the LAI and annual precipitation near 16°N are also plotted in Figs. 7(a) and (b) respectively, labelled as Exp5. Although the climate in the year of irrigation is significantly wetter and greener, it gradually decays back to the control state. Interest-ingly, further increase of grass density fails to cause the transition of the biosphere–atmosphere system from the control equilibrium to a wetter one. The control climate presented in Fig. 5 might be the most favourable equilibrium the climate system can maintain within the perturbation regime of this study, i.e. removal or enhancement of grass between 12.5°N and 17.5°N.

In the experiments presented above, vegetation degradation over the grassland region caused climate transitions from the control equilibrium to drier equilibria. These transitions are reversible. Further experiments demonstrated that, for the climate system at a drier equilibrium, favourable perturbations such as irrigation can trigger the system to develop into a wetter equilibrium. As an example, Fig. 8 presents the change of the annual rainfall at 16°N after an irrigation effect is imposed on the biosphere–atmosphere
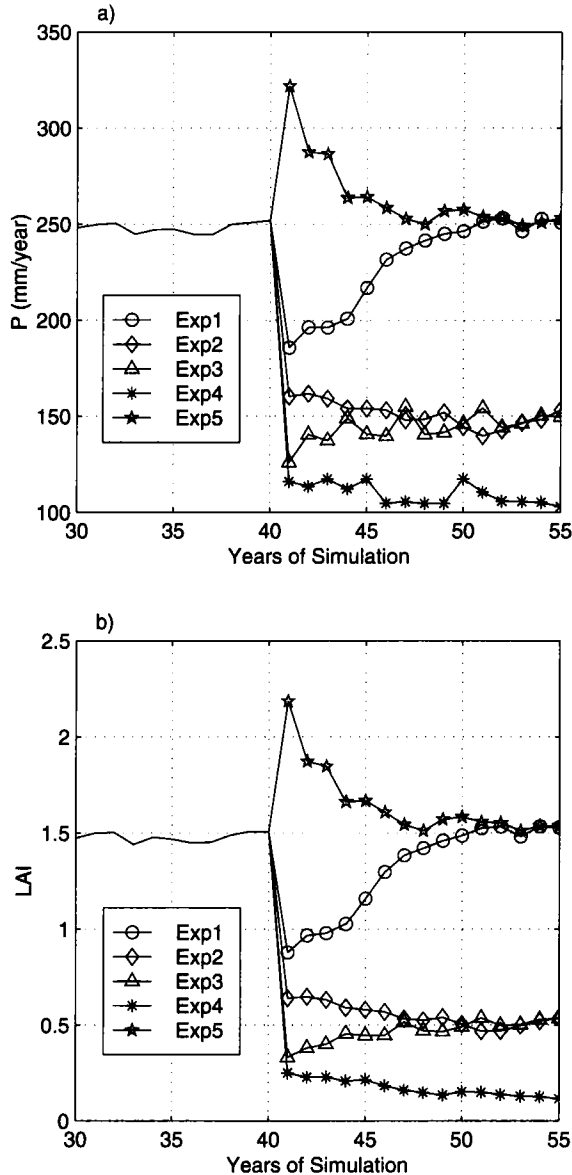
Figure 7. Evolution of the biosphere–atmosphere system before and after various perturbations (Exp1–Exp5, see text) that take place in the 41st year over the grassland region: (a) precipitation at 16°N, and (b) growing-season leaf area index for grass at 16°N.

system at the dry equilibrium New1. Here the irrigation features a 30% increase of grass density, which is a change of small magnitude given the low vegetation density at the grassland region of equilibrium New1. As shown in Fig. 8, the irrigation perturbation induces a positive feedback which eventually leads the system to the wet equilibrium.

For a complete understanding of the climate system over West Africa, in the following we look at additional characteristics of the new climate equilibria in comparison with the control equilibrium. The normalized difference in annual precipitation and NPP between the control and the new equilibria are shown in Figs. 9(a) and (b), respectively. Although very little difference is observed over the forest region, both of the new
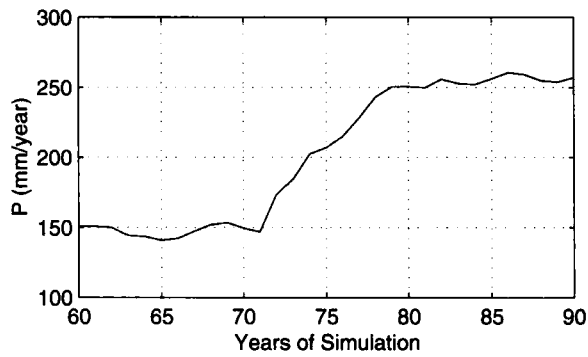
Figure 8. The evolutionary pattern for precipitation at 16°N when irrigation is introduced in the biosphere–atmosphere system at the dry equilibrium New1 (see text) during the 72nd year of the simulation.
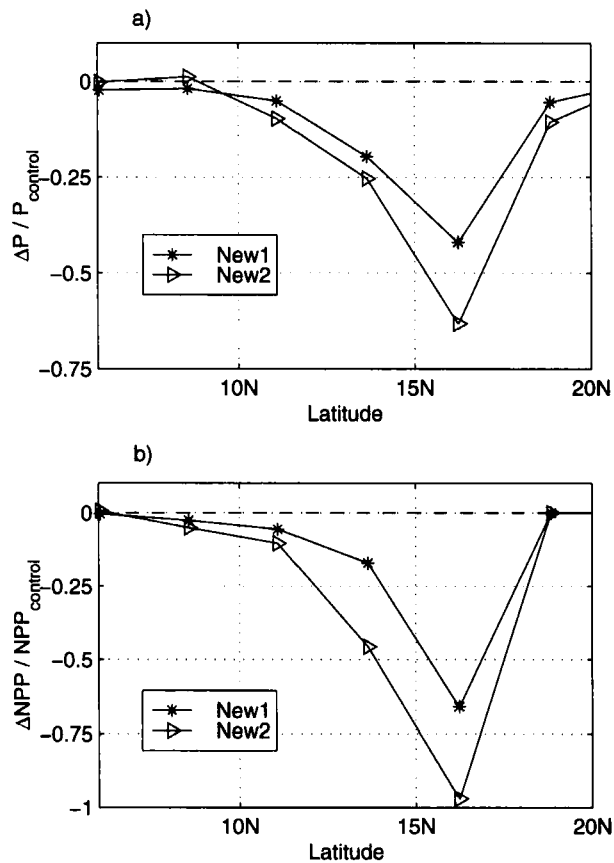

Figure 9. The normalized difference between the control equilibrium and the new equilibria (New1 and New2), 'New'−Control: (a) precipitation, (b) net primary productivity.
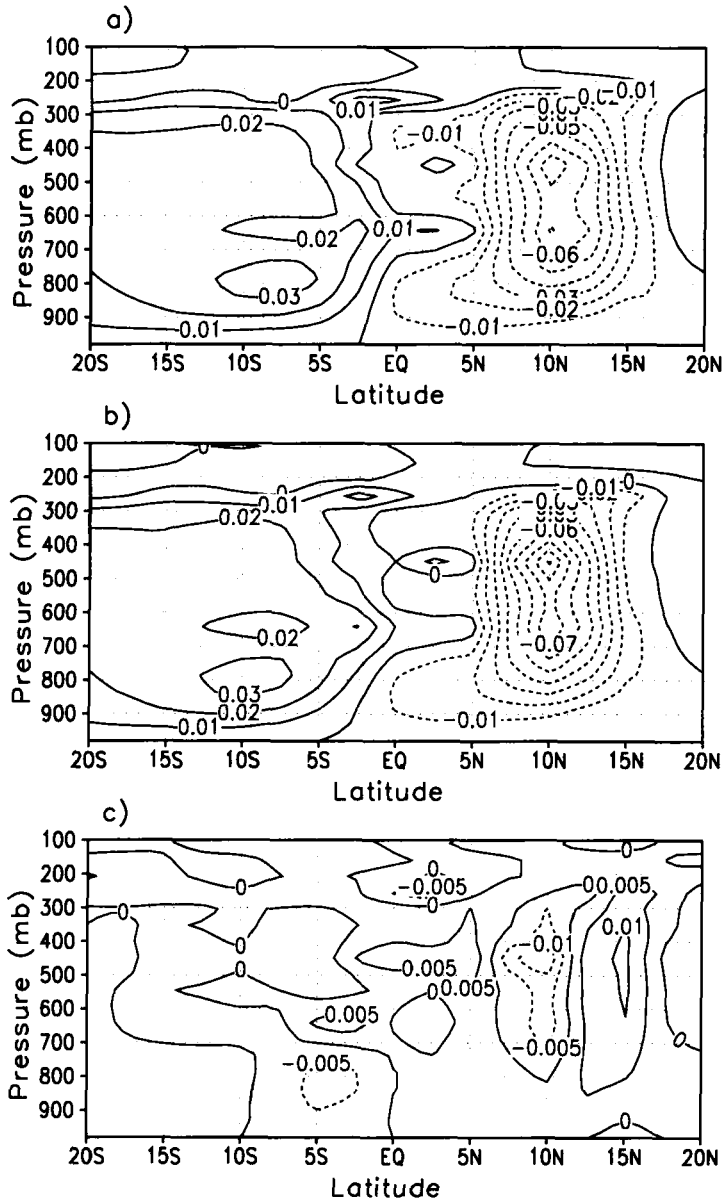
Figure 10.   The vertical velocity (Pa s^{-1}) in August: (a) the control equilibrium, (b) the New2 equilibrium, and (c) the difference between New2 and the control (New2−control). Negative contours are dashed.

equilibria are significantly drier and less productive than the control equilibrium over the grassland region. Correspondingly, over the grassland region and during the rainy season, the local ascending motion of the new equilibria is weaker, and this difference can be as much as 50% around 15°N in August for the equilibrium New2 (Fig. 10). Compared with the control equilibrium, the rising branch for equilibrium New2 features a southward shrink over the Sahel region. However, there is no overall southward shift of the monsoon circulation. The enhancement of the ascent observed in Fig. 10(c) is limited
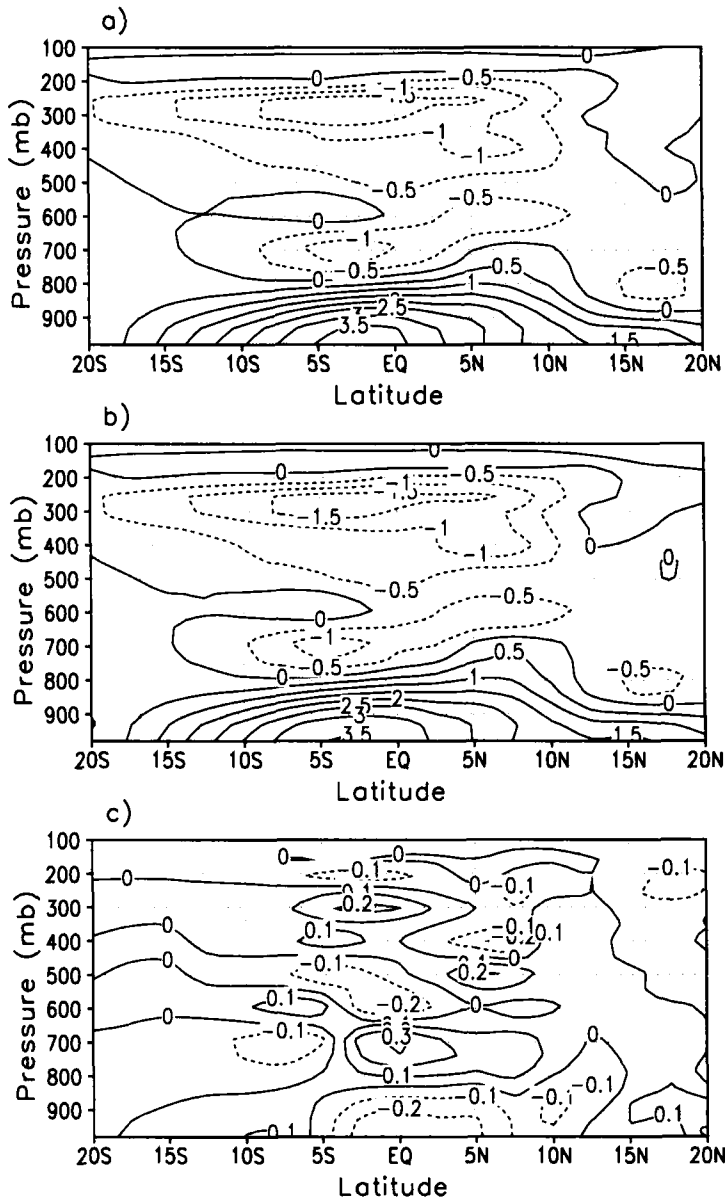
Figure 11.   As Fig. 10 for the meridional wind (m s^{-1}).

to the centre of the rising branch. Figure 11 presents a comparison of the meridional-wind field in August between the control and the New2 equilibrium. At low levels, New2 features weaker northward wind across the coast, which signals a slightly weaker monsoon circulation. Simultaneously, over the region 12°N–16°N, the pattern of the change in meridional wind (Fig. 11(c)) resembles that of a low-level divergence (negative in the south and positive in the north), which implies weaker local convection. Therefore, the arid condition at the new equilibria over the grassland region is likely to be associated with both a weaker large-scale circulation and suppressed local convection.

Figure 12.   The surface temperature difference between the control equilibrium and the two new equilibria
('New'—Control) during: (a) August (rainy season), and (b) February (dry season).

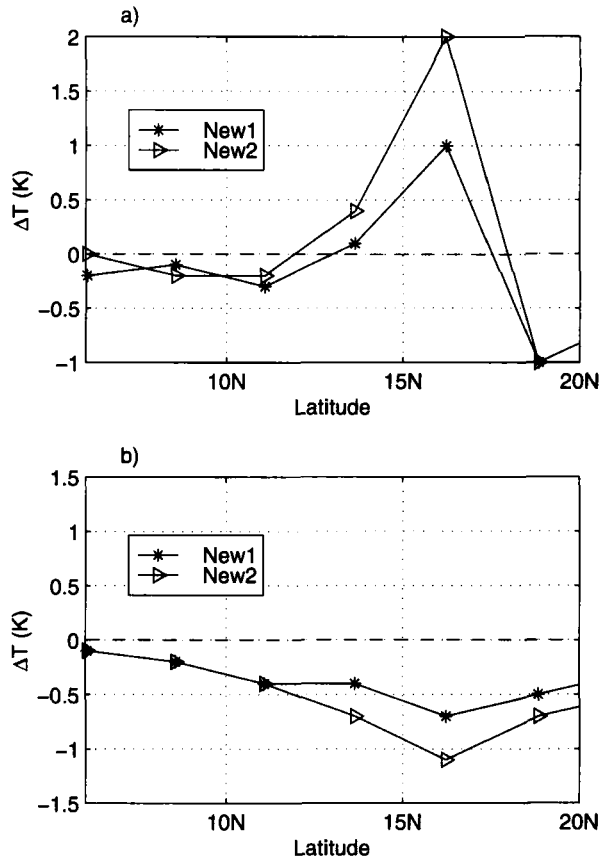Comparison of the surface temperature between different equilibria shows an inter-
esting seasonal pattern. At the grassland region, the new equilibria are warmer during
the rainy season but cooler during the dry season, as shown in Figs. 12(a) and (b) which
presents August and February as examples. The warming in the rainy season takes place
in spite of a decrease in the net radiation (Fig. 13(a)). Upon vegetation degradation,
albedo increase reduces the net radiation, which tends to cool the land surface (Charney
1975); on the other hand, evapotranspiration decreases, which tends to warm up the land
surface (Idso 1977). Figure 13(b) presents the difference between the new equilibria and
the control equilibrium in the net solar radiation of August, which is associated with
the albedo effect; Fig. 13(c) presents the difference in the latent-heat flux of August,
which is associated with the evapotranspiration effect. Clearly, the evapotranspiration
effect is dominant during the rainy season over the grassland region, thus causing the
local increase of surface temperature observed in Fig. 12(a). However, the albedo effect
becomes dominant during the dry season when the evapotranspiration is minimal.

(c)   *Summary*

The above experiments on different types of vegetation perturbations imply that
the simulated biosphere–atmosphere system is more resistant to perturbations over the
forest region than to perturbations over the grassland region. If the ecosystem is saved
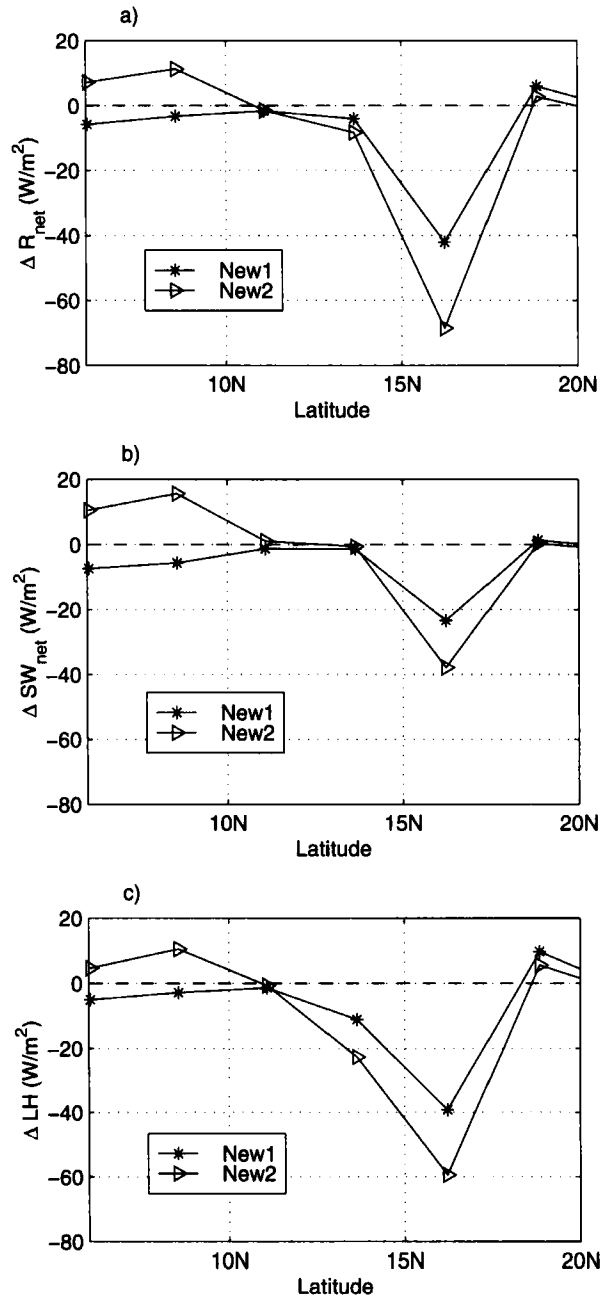
Figure 13. The difference in surface energy fluxes between the control equilibrium and the new equilibria in August ('New'−Control): (a) net radiation, (b) net solar radiation, and (c) latent-heat flux. The decrease of the net solar radiation tends to cool the land surface, while the decrease of the latent-heat flux tends to warm up the land surface.

from further human disturbances, the biosphere–atmosphere system can recover from a forest disturbance, while it may not recover from a grassland disturbance. However, this does not imply that deforestation is less significant than desertification. In fact, as shown in Fig. 6, the deforestation-induced decrease in precipitation is quite significant during the early stage of the succession which lasts for several decades.

As demonstrated clearly by Figs. 7(a) and (b), experiments Exp1, Exp2, and Exp3 represent three different types of responses by the coupled biosphere–atmosphere system: a negative feedback leading to a full recovery (Exp1); a positive feedback leading to perturbation enhancement (Exp2); and a negative feedback leading to a partial recovery (Exp3). The response of the climate system in experiment Exp4 is of the same type as Exp2, but within the attraction zone of a different equilibrium. Our results from the ZonalBAM support the theoretical hypothesis of section 2 on the necessary conditions for the existence of multiple equilibria and on how the coupled biosphere–atmosphere system responds to non-permanent vegetation perturbations.

5. Discussion

Although in this study we focus on anthropogenic vegetation perturbations, natural climate variations can cause similar effects in degrading the vegetation, especially in the grassland region. An event of one extremely dry year or several dry years in a row could cause as much vegetation degradation as man-made desertification. Similarly, a naturally induced wet event can significantly enhance the grass growth. Therefore, transitions between different equilibria shown in Fig. 7 do not depend on human activities. Historically, and before the emergence of human activity as a significant process, the main forcing in triggering a climate transition would be the internal climate variations and other natural factors, e.g. lightning-ignited fires, plant pathogens, and diseases of grazing fauna.

The existence of multiple climate equilibria, and, equally important, the possibility of climate transitions between different equilibria, have significant implications regarding the past, present and future climate over West Africa. The alternate occurrences of dry and wet spells over West Africa (see Fig. 1) can be viewed as a collective reflection of the climate persistence at one equilibrium and the climate transition to another, triggered by external forcings and governed by the two-way biosphere–atmosphere feedback including ecosystem dynamics. For the current drought, it is conceivable that the climate system before the drought might have been in a state similar to the control equilibrium of this study. The two new equilibria explored in this study, New1 and New2, are both significantly more arid than the control equilibrium (Fig. 9). A single dry year or several dry years in a row, or man-made desertification, might have triggered the climate system to evolve into a drier equilibrium. An event of one or more significantly wet years in the future, or man-made perturbations such as large-scale irrigation, may trigger the system towards a wetter equilibrium thus starting a wet period. The irrigation-induced recovery of the wet equilibrium, as shown in Fig. 8, brings up the possibility of mitigating the current drought by way of human perturbations. A 3-D model with a finer resolution is required for an accurate estimation on how much irrigation is needed to induce such a beneficial effect.

In the above interpretation of climate variability over West Africa, we do not intend to exclude the impact of other factors such as the ocean dynamics. In fact, a large number of previous studies (Lamb 1978a,b; Hastenrath 1984; Lough 1986; Folland *et al.* 1991; Rowell *et al.* 1995; Zheng 1998; Zheng *et al.* 1999) provided evidence that sea surface temperature in the Atlantic ocean may influence rainfall in the Sahel region. Without

incorporating the oceanic processes, it is impossible to address whether the biosphere–atmosphere feedback is the most significant mechanism in causing the observed long-term climate variations. Our argument here is rather that the biosphere–atmosphere feedback alone can act as a significant mechanism regulating the low-frequency climate variability over that region. The important role of the ocean is the focus of some of our on-going research.

The multiple-equilibrium nature of the climate system adds more uncertainty and challenge to climate studies using coupled biosphere–atmosphere models (e.g. Foley *et al.* 1998). Discrepancies between model simulations and observations may not necessarily reflect a deficiency in the model, since observations and simulations may possibly describe two different equilibria. It is technically difficult and computationally expensive to identify all the relevant equilibria that the model might have, especially for 3-D models.

Finally, our results have important implications regarding the general topic of climate predictability. The perturbation-induced climate transition between different equilibria implies that climate in West Africa should be considered as an initial-value problem (Pielke 1998) as well as a boundary-value problem. The traditional notion of climate as exclusively a boundary-value problem, which is used to justify most of the current approaches for predicting future climate, is seriously challenged. Our finding necessitates a re-evaluation of the current understanding regarding climate predictability, and calls for new approaches to climate prediction.

6. CONCLUSIONS

This paper focuses on the role of ecosystem dynamics in biosphere–atmosphere interactions over West Africa. We present both theoretical hypotheses and modelling results on how the synchronously coupled biosphere–atmosphere system responds to vegetation perturbations. The main conclusions include:

(i) The regional biosphere–atmosphere system over West Africa has multiple equilibrium states coexisting under the same external orbital forcing.

(ii) Upon certain perturbations, the biosphere–atmosphere system can evolve from one equilibrium to another within a relatively short time (on the order of one decade).

(iii) Following a perturbation, the biosphere–atmosphere system has three different ways of responding: a negative feedback leading to a full recovery; a positive feedback leading to perturbation enhancement (therefore a new equilibrium); a negative feedback leading to a partial recovery (therefore a new equilibrium).

The scope of this paper is limited to the response of the biosphere–atmosphere system to a major disturbance. In reality, the climate system is constantly subject to natural variability in the atmospheric and oceanic processes. Based on the understanding developed in this study, further research on how the coupled biosphere–atmosphere system responds to continuous natural variations (e.g. the seasonal and inter-annual variability in the Atlantic SST) is underway.

While previous studies looked at the climatic impact of permanent land-surface changes, this study focuses on non-permanent vegetation perturbations and aims at understanding the underlying natural processes. Except for the one-time perturbation, the ecosystem processes in our model are all natural. However, today's ecosystem is often a combination of both natural and managed ones. While natural successions always take place, recurrent perturbations may prevent part of the landscape from recovering fully, e.g. fire is frequently used to maintain some deforested region as

pasture. As a result, the landscape is subject to a mosaic combination of permanent and non-permanent perturbations. Future research will look at this regional climate system with an ecosystem model that combines both the natural and managed processes.

Although we focus here on the region of West Africa, lessons of this study have general ecological implications. Today, the degradation of ecosystems is worldwide, the intensity of vegetation loss and the expansion of land use are striking. It is of importance for ecologists, economic developers, and policymakers to bear in mind that the resilience of the ecosystem is not unlimited. Intense exploitation can lead the environment into a self-degrading phase. The need for preservation can never be overemphasized.

REFERENCES

Charney, J. G. 1975 Dynamics of deserts and drought in the Sahel. *Q. J. R. Meteorol. Soc.*, **101**, 193–202

Charney, J. G., Quirk, W. J., Chow, S. and Kornfield, J. 1977 A comparative study of the effects of albedo change on drought in semi-arid regions. *J. Atmos. Sci.*, **34**, 1366–1385

Claussen, M. 1997 Modeling bio-geophysical feedback in the African and Indian monsoon region. *Clim. Dyn.*, **13**, 247–257

1998 On multiple solutions of the atmosphere–vegetation system in present-day climate. *Global Change Biology*, **4**, 549–559

Demaree, G. R. and Nicolis, C. 1990 Onset of Sahelian drought viewed as a fluctuation-induced transition. *Q. J. R. Meteorol. Soc.*, **116**, 221–238

Eltahir, E. A. B. 1996 Role of vegetation in sustaining large-scale atmospheric circulations in the tropics. *J. Geophys. Res.*, **101**, 4255–4268

Eltahir, E. A. B. and Gong, C. 1996 Dynamics of wet and dry years in West Africa. *J. Clim.*, **9**, 1030–1042

Entekhabi, D., Rodriguez-Iturbe, I. and Bras, R. L. 1992 Variability in large-scale water balance with land surface–atmosphere interaction. *J. Clim.*, **5**, 798–813

Ewel, J. 1971 Biomass changes in early tropical succession. *Turrialba*, **21**, 110–112

1983 Succession. Pp. 217-223 in *Ecosystems of the World 14A*. Ed. F. B. Golley. Elsevier, New York

Farmer, G. and Wigley, T. M. L. 1985 Climatic trends for tropical Africa. Research Report, University of East Anglia, Norwich, UK

Foley, J. A., Levis, S., Prentice, I. C., Pollard, D. and Thompson, S. L. 1998 Coupling dynamic models of climate and vegetation. *Global Change Biology*, **4**, 561–579

Folland, C., Owen, J., Ward, M. N. and Colman, A. 1991 Prediction of seasonal rainfall in the Sahel region using empirical and dynamical methods. *J. Forecasting*, **10**, 21–56

Golley, F. B. 1975 'Productivity and mineral cycling in tropical forests'. Pp. 106–115 in Proceedings of a symposium presented at the fifth general assembly of the special committee for the I.B.P. National Academy of Science, Washington, DC

Golley, F. B., McGinnis, J. T., Clements, R. G., Child, G. I. and Duever, M. J. 1975 *Mineral cycling in a tropical moist forest ecosystem.* University of Georgia Press, Athens, Ga

Hastenrath, S. L.   1984   Interannual variability and annual cycle: Mechanisms of circula-
                                          tion and climate in the tropical Atlantic sector. *Mon. Weather
                                          Rev.*, **112,** 1097–1107

Idso, S. B.   1977   Note on some recently proposed mechanisms of genesis of
                                          deserts. *Q. J. R. Meteorol. Soc.,* **103,** 369–370

Lamb, P. J.   1978a   Case studies of tropical Atlantic surface circulation patterns dur-
                                          ing recent sub-Saharan weather anomalies: 1967 and 1968.
                                          Mon. Weather Rev., **106,** 482–491

                                  1978b   Large-scale tropical Atlantic surface circulation patterns associ-
                                          ated with sub-Saharan weather anomalies. *Tellus,* **30,** 240–
                                          251

Lough, J. M.   1986   Tropical Atlantic sea surface temperatures and rainfall variations
                                          in sub-Saharan Africa. *Mon. Weather Rev.,* **114,** 561–570

Maley, J.   1973   Mecanisme des changements climatiques aux basses latitudes.
                                          Palae. Paleo. Paleo., **14,** 193–227

                                  1981   Etudes palynologiques dans le bassin du Tchad et
                                          pal'eoclimatologie de l'Afrique nord-tropicale de 30 000 ans
                                          a l'epoque actuelle. Technical Report No. 129, l'ORSTOM,
                                          Paris, France

Nicholson, S. E.   1981   The historical climatology of Africa. Pp. 249–270 in *Climate and
                                          history*. Eds. T. M. L. Wigley *et al.* Cambridge University
                                          Press, Cambridge

                                  1989   African drought: Characteristics, causal theories and global tele-
                                          connections. *Geophysical Monograph,* **52,** 79–100

Nicolis, C.   1982   Stochastic aspects of climate transitions—response to a periodic
                                          forcing. *Tellus,* **34,** 1–9

Pielke, R. A.   1998   Climate prediction as an initial value problem. *Bull. Am.
                                          Meteorol. Soc.,* **79,** 2743–2746

Reynolds, R. W. and Smith, T. M. 1994   Improved global sea surface temperature analysis using optimum
                                          interpolation. *J. Clim.,* **7,** 927–948

Rodriguez-Iturbe, I., Entekhabi, D. 1991 Nonlinear dynamics of soil moisture at climate scales. 1: Stochas-
    and Bras, R. L.   tic analysis. *Water Resour. Res.,* **27,** 1899–1906

Rowell, D. P., Folland, C. K.,   1995   Variability of summer rainfall over tropical north Africa (1906–
    Maskell, K. and Ward, M. N.   92): Observations and modelling. *Q. J. R. Meteorol. Soc.,*
                                          121, 669–704

Sud, Y. C. and Molod, A.   1988   A GCM simulation study of the influence of Saharan evapo-
                                          transpiration and surface-albedo anomalies on July circula-
                                          tion and rainfall. *Mon. Weather Rev.,* **116,** 2388–2400

Wang, G. and Eltahir, E. A. B.   2000   Biosphere–atmosphere interactions over West Africa. 1: Devel-
                                          opment and validation of a coupled dynamic model. *Q. J. R.
                                          Meteorol. Soc.,* **126,** 1239–1260

Wang, J., Bras, R. L. and   1997   Structure in fluctuations of large-scale soil moisture climate due
    Entelshabi, D.   to external random forcing and internal feedbacks. *Stochastic
                                          Hydrology and Hydraulics,* **11,** 95–114

Xue, Y. and Shukla, J.   1993   The influence of land surface properties on Sahel climate. Part I:
                                          Desertification. *J. Clim.,* **6,** 2232–2245

Zhang, H. and   1996   Impacts of tropical deforestation. Part I: Process analysis of local
    Henderson-Sellers, A.   climate change. *J. Clim.,* **9,** 1497–1517

Zheng, X.   1998   The response of a moist zonally symmetric atmosphere to sub-
                                          tropical surface temperature perturbation. *Q. J. R. Meteorol.
                                          Soc.,* **124,** 1209–1226

Zheng, X. and Eltahir, E. A. B.   1997   The response to deforestation and desertification in a model of
                                          West African monsoons. *Geophys. Res. Lett.,* **24,** 155–158

Zheng, X., Eltahir, E. A. B. and 1999   A mechanism relating tropical Atlantic spring sea surface temper-
    Emanuel, K. A.   ature and West African rainfall. *Q. J. R. Meteorol. Soc.,* **125,**
                                          1129–1164