Global Change Biology (1998) **4**, 461–475

Interactions between the atmosphere and terrestrial ecosystems: influence on weather and climate

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Abstract

This paper overviews the short-term (biophysical) and long-term (out to around 100 year timescales; biogeochemical and biogeographical) influences of the land surface on weather and climate. From our review of the literature, the evidence is convincing that terrestrial ecosystem dynamics on these timescales significantly influence atmospheric processes. In studies of past and possible future climate change, terrestrial ecosystem dynamics are as important as changes in atmospheric dynamics and composition, ocean circulation, ice sheet extent, and orbit perturbations.

Keywords: atmospheric and ecosystem dynamics, biogeophysical/biogeochemical interactions with weather and climate, climate, climate change, global change, land–atmosphere interaction

Received 28 March 1997; revised version received 12 September and accepted 14 October 1997

Introduction

Terrestrial ecosystem–atmospheric interactions refer to exchanges of heat, moisture, trace gases, aerosols, and momentum between land surfaces and the overlying air. These feedbacks represent a dynamic coupled system which would be expected to evolve differently as a result of the interactions between the two media. This article summarizes these interactions and discusses why they are important.

Terrestrial ecosystems and climate influence one another on timescales ranging from seconds to millions of years (e.g. see fig. 1 in Sellers *et al*. 1995). Ecosystems influence weather and climate over period of seconds to years through exchanges of energy, moisture, and momentum between the land surface and the atmosphere (Waggoner & Reifsnyder 1968; Shukla & Mintz 1982; Dickinson 1984; Dickinson *et al*. 1986; Sellers *et al*. 1986; Pan & Mahrt 1987; Dickinson *et al*. 1992; Bonan 1994; Sellers *et al*. 1996a) and the changes in global-scale atmospheric circulation that can result from changes in these fluxes (Charney 1975; Chervin 1979; Sud & Fennessy 1982; Sud & Smith 1985; Sud *et al*. 1988; Dickinson & Henderson-Sellers 1988; Delworth & Manabe 1989; Nobre *et al*. 1991; Bonan *et al*. 1992; Franchito & Rao 1992; McGuffie *et al*. 1995; Chase *et al*. 1996; Randall *et al*. 1996; Sellers *et al*. 1996c; Zhang *et al*. 1996). Ecosystem structure and function is strongly determined on timescales of decades to centuries by climate influences, primarily through temperature ranges and water availability (Woodward 1987; Woodward & McKee 1991; Prentice *et al*. 1992; Neilson & Marks 1994; Neilson 1995; Henderson-Sellers & McGuffie 1995; Bugmann 1997; Chapin & Starfield 1997). On timescales of thousands of years, glacial–interglacial cycles probably involve coupled changes in the geographical distribution of terrestrial ecosystems, surface albedo, biogeochemical cycling, and climate (e.g. see Foley 1994; Foley *et al*. 1994; Dutton & Barron 1996, 1997; Prentice *et al*. 1993; Gallimore & Kutzbach 1996; Crowley & Baum 1997; Adams *et al*. 1990; Harrison *et al*. 1995) in response to changing solar forcing. On even longer geological timescales (millions of years), terrestrial ecosystems and the Earth's climate have evolved together through such mechanisms as changes in the biochemistry and the composition of the atmosphere (Sagan & Mullen 1972; Budyko 1977; Worsley & Nance 1989; Kasting 1993).

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Our paper concentrates on atmosphere-terrestrial ecosystem interactions on timescales of a hundred years and less. For longer time periods, the interactions become even more intertwined as slower time period processes (e.g. continental glaciation) become important. To adequately understand those slower time period effects, however, necessitates we understand the more rapid interactions (100 years and less) which is the focus of our paper.

Modelling of meteorological flows requires the use of conservation equations for fluid velocity, heat, mass of dry air, water substance in its three phases, and many other natural and anthropogenic atmospheric constituents. Spatial scales of simulation have ranged from high resolution representations of the boundary layer where grid increments are on the order of tens of meters or less, to general circulation representations of the entire globe.

The characterization of biospheric processes in these models, however, has been limited to simple representations where most aspects of the soil and vegetation are prescribed. Stomatal conductance responds to atmospheric inputs of solar radiation, air temperature, air relative humidity, precipitation, and air carbon dioxide concentration, and to soil temperature and moisture. Up to the present, in meteorological models, these are the only meteorological variables to which the vegetation and the soil dynamically respond.

Recent examples of model investigation of the importance of land surface conditions on weather and climate include: Otterman & Chou (1992); Ek & Cuenca (1994); Keller & Matson (1994); Panajiston (1995); Xinmei *et al.* (1995a); Ye & Xinyuan (1995); Avissar & Liu (1996); Betts *et al.* (1996); Crook (1996); Cuenca *et al.* (1996); Dirmeyer & Shukla (1996); Giorgi *et al.* (1996); Koster & Suarez (1996); Kutzbach *et al.* (1996); Levis *et al.* (1996); Luthi *et al.* (1996); Lyons *et al.* (1996); Molders & Raabe (1996); Pan *et al.* (1996); Qu *et al.* (1996); Rabin & Martin (1996); von Salzen *et al.* (1996); Xue (1996); Zhang *et al.* (1996); Crowley & Baum (1997); Pielke *et al.* (1997); Schar *et al.* (1997); Bornstein *et al.* (1994).

Recent observationally based papers on this topic include: Lyons *et al.* (1993); Xinmei *et al.* (1995b); Baldocchi & Shankar Rao (1996); Da Rocha *et al.* (1996); Grunwald *et al.* (1996); Jones (1996); Livezey & Tinker (1996); Nemani *et al.* (1996); Prueger *et al.* (1996); and Avissar *et al.* (1998) A special issue of the *Journal of Geophysical Research* (20 March 1996) edited by R. Avissar and M. Coughlan, included a wide range of papers on this topic (e.g. Avissar & Liu 1996; Betts *et al.* 1996; Chase *et al.* 1996; Chen *et al.* 1996; Copeland *et al.* 1996; Cuenca *et al.* 1996; Gao *et al.* 1996; Rabin & Martin 1996; Seth & Giorgi 1996; Shao & Henderson-Sellers 1996; Wetzel *et al.* 1996; Xue *et al.* 1996). (A listing of other published work in these research areas can be found in Dalu *et al.* (1996), Graetz (1991), and Williams & Balling (1996).)

The modelling of terrestrial ecosystems involves the short-term response of vegetation and soils to atmospheric effects, and the longer-term evolution of species composition, biome dynamics, and nutrient cycling associated with landscape and soil structure changes. The assimilation of carbon resulting in the growth of vegetation, and its subsequent release during decay has been a focus of these models (e.g. see Parton *et al.* 1987, 1988; Running 1994; Hunt *et al.* 1996; Parton 1996). The spatial scales of these simulations have ranges from patch sizes (microscale) to biome (mesoscale) scales. These models often include a stochastic component to represent unpredictable random inputs from the atmosphere and interactions within the vegetation such as fires, for example.

These models require such atmospheric inputs as temperature, relative humidity, net radiation, and precipitation in order to integrate their formulations forward in time. Output from nearby climatological stations have been used as the needed boundary conditions for these models when applied on the patch up to the regional scale. On the global scale, outputs from general circulation models have been used to estimate potential changes of biome type in response to hypothesized climate change scenarios using, for example, the concept of a *Holdridge* diagram (Henderson-Sellers & McGuffie 1995) and other vegetation models (Claussen 1991; Prentice *et al.* 1992).

Short-term interactions

The vertical structure of the daytime atmospheric boundary layer is critically dependent on the partitioning of net radiation into sensible and latent turbulent heat flux, and groundheat conduction. A deeper boundary layer, for example, results when more of this radiative energy is realized as sensible heat flux. This results because sensible heat flux generates buoyancy at the land surface, which increases the turbulent kinetic energy of the convective boundary layer (CBL), allowing it to grow by turbulent entrainment against static stability at the CBL top. When vegetation is present, the response of leaf conductance to atmospheric conditions represents a rapid feedback between the biosphere and the atmosphere. The passage of a cloud during daylight, for example, will significantly reduce short-wave radiation with stomata apertures responding within minutes.

The drying of the near surface soil, and the depletion of deeper soil moisture as a result of transpiration represent another, somewhat slower feedback with the atmosphere which varies over days to a few weeks. When vegetation becomes water stressed, for instance, stomata will close in order to conserve the remaining water, so that a larger fraction of the net radiation is realized as sensible heat flux (Avissar & Pielke 1991). Precipitation represents a short-term feedback that can quickly replenish the soil moisture, as well as provide shallow liquid water layers on the vegetation (interception).

Long-term interactions

Seasonal interactions include feedbacks between increases in LAI from spring into the summer which will modify the partitioning of latent and sensible heat fluxes. Nutrient limitations can also constrain biomass growth, particularly in moist environments. It has been shown that the increase of mean temperature in the spring in the eastern United States is interrupted as vegetation leafs out due to the shift in the surface energy budget from sensible heating to latent heating (Schwartz & Karl 1990). Additionally, drought conditions over the eastern United States are apparently perpetuated by the reduced transpiration from the water-stressed vegetation.

Other important influences of the land surface on the atmosphere include the modification of the albedo of the surface. The albedo has been shown to be a critical landscape–atmospheric interaction where, for example, desertification of the Sahel region of Africa may have resulted from excessive grazing by domesticated goats of the darker vegetation such that a larger fraction of the solar radiation is reflected back into space (the field campaign HAPEX-Sahel was expected to provide data to understand this effect—Dolman *et al.* 1997; Goutorbe *et al.* 1997). It has been suggested that the Rajastan desert in India has resulted from the same mechanism.

On time periods of years, species composition and soil characteristics including nutrient turnover can change in response to long-term atmospheric changes, through natural vegetation succession, human disturbances, and episodic events such as fire and severe windstorms. These terrestrial-ecosystem changes would be expected to then feedback to the atmospheric structure. Obvious prehistoric examples of these changes include the dynamic changes of the landscape as boreal and temperate biome regions shifted poleward after the retreat of the maximum Pleistocene glaciation.

Field programmes

Recently, there have been several field programmes which focus on land surface–atmosphere interactions such as: FIFE (Sellers *et al.* 1992a, Betts & Beljaars 1993, Smith 1997) in Kansas, USA; BOREAS (see Sellers *et al.* 1995) in Saskatchewan and Manitoba, Canada; LOTREX (Schadler *et al.* 1990) in Hildesheimer Borde, Germany; EFEDA (Bolle *et al.* 1993) in Spain; HAPEX-MOBILHY (Andre *et al.* 1989a; Noilhan *et al.* 1991) in France; HAPEX-SAHEL (e.g. Goutorbe *et al.* 1994, 1997) in Niger; SEBEX (Wallace *et al.* 1991) in the Sahel of Africa; and in the Amazon region (e.g. Shuttleworth 1985; Wright *et al.* 1992; Gash & Shuttleworth 1991; Nobre *et al.* 1996; Shuttleworth *et al.* 1991); and REKLIP in central Europa (Parlow 1996). The Global Energy and Water Cycle Experiment (GEWEX) will provide additional data to assess land-surface–atmosphere interactions (Coughlan & Avissar 1996).

Human dimensions

The socioeconomic and political involvement of humans also needs to be factored into atmosphere–land surface interactions. Martin *et al.* (1996) provides a discussion of this issue with respect to desertification. Raup (1957) describes how human disturbance prevents vegetation composition from achieving any line of 'climax' or 'equilibrium' state. This thesis is further elaborated on, and confirmed in the seminal study of New England landscape by Foster (1992, 1993, 1995) and Foster *et al.* (1992). Motzkin *et al.* (1996) concluded that 'The widespread and long-lasting impact of human activity on natural systems indicates that land-use history must be treated as an integral aspect of ecological study . . .'. Kittel *et al.* (1996) present a procedure to collect land-surface data for use in land-atmosphere studies. A book entitled *The Earth as Transformed by Human Action*, edited by Turner *et al.* (1990) includes several papers directly relevant to this topic (e.g. Richards 1990; Williams 1990).

Land–atmosphere carbon change

Although land covers only 29% of the Earth's surface, and not all land is vegetated, the storage of organic carbon in terrestrial ecosystems, which is 1600 Gt C (1 Gt C = 10^9 kg of carbon) is more than twice the storage of carbon in the atmosphere (765 Gt C) as CO_2. Annual exchanges of carbon between these reservoirs amounts to about 120 Gt C, with creation of organic matter by photosynthesis approximately balanced by release of CO_2 by respiration and decomposition (Schimel *et al.* 1996). By contrast, the global emissions of CO_2 due to combustion of fossil fuels is only about 6 Gt C (Andres *et al.* 1996), so a perturbation in the natural carbon exchange of terrestrial ecosystems of only 5% would be equal to the total industrial emission to the atmosphere. Terrestrial biomass is being destroyed by deforestation, biomass burning, and other land use changes at a rate of between 0.6 Gt C y^{-1} and 2.0 Gt C y^{-1}, with the majority of this flux occurring in the tropics (Houghton *et al.* 1987, Houghton 1993). Nevertheless, it appears that the overall contribution of the terrestrial biosphere to the carbon balance of the atmosphere is to act as a sink of 1–2 Gt C y^{-1} (Schimel *et al.* 1996). This may be due to fertilization by CO_2 itself (Gifford 1994; Friedlingstein *et al.* 1995), nitrogen deposition (Schindler & Bayley 1993); climate fluctuations (Dai & Fung 1993), reforestation (Dixon *et al.* 1994), or some combinations of these and other unknown factors. The exchanges of carbon between the atmosphere and land surface are quite sensitive to changes in climate forcing such as temperature, precipitation, and radiation, and in turn have the potential to significantly affect the radiative balance of the atmosphere through changes in the mixing ratio of CO_2, if systematic perturbations persist for years to decades. In addition, CO_2 concentration has a direct effect on plant stomatal physiology by which the surface energy balance is intimately linked to atmospheric CO_2 (Collatz *et al.* 1991, 1992), although some studies have found this linkage to be weaker in the field than in the lab (Korner & Arnone 1992). This physiological response may in turn have regional and global implications for atmospheric circulation patterns and climate (Sellers *et al.* 1996c; see Section 3).

Mechanisms influencing biosphere–atmosphere feedbacks

The surface energy budget

The surface energy budget (SEB) of a landscape is governed by an interacting set of surface and atmospheric processes which operate at timescales from minutes to many years. At timescales of days to years, the SEB is linked with the water, carbon and nutrient balances of the surface, and thence to ecosystem development. One key linkage of this type occurs through evaporation (both from the soil and as plant transpiration) which provides the common term between the SEB and the soil water balance. A second group of SEB–carbon–nutrient linkages occurs through the land–air carbon flux, via the effects of light and water on plant growth. These critical resources are constrained by SEB processes such as the radiation balance and latent heat flux, yet in turn they largely determine the vegetation state both above ground (influencing albedo and transpiration, for instance) and below ground (influencing rooting distribution and thus the capacity of the vegetation to gather soil resources, e.g. see Nepstad *et al.* 1994; Desborough 1997; Koster & Milly 1997). A third set of links occurs through human-induced landcover changes, including clearing, herbivory and fire. All of these processes in turn feed back onto atmospheric behaviour, largely through the SEB.

Other important biosphere–atmosphere feedbacks act on the SEB over much shorter (subdiurnal) timescales, and are therefore essentially instantaneous from an ecosystem viewpoint. It has long been recognized that the short-term radiative, aerodynamic, physiological and boundary-layer processes determining the SEB are not independent (Thom 1975; McNaughton & Jarvis 1991). Another paper in this issue (Raupach 1998, this issue) quantifies the interactions between four short-term feedback pathways, all of which can be considered as acting on the surface temperatures of the vegetation and soil. These can be very different, particularly when the vegetation is sparse, but appropriate averaging procedures can be used to define a bulk surface temperature T_s (e.g. Raupach & Finnigan 1995). The four feedbacks evaluated by Raupach (1998) are:

1 *Radiative coupling* The available energy for transfer to the atmosphere as sensible and latent heat is modulated by the effect of T_s on outgoing longwave radiation (Monteith 1973). This feedback is small except over smooth, dry surfaces, and is negative in daytime conditions because a positive T_s perturbation decreases the available energy by increasing the outward longwave radiation, thereby inducing a restoring T_s perturbation.

2 *Aerodynamic coupling* The turbulent transfer of heat and moisture is modulated by atmospheric stability and therefore by the SEB, feeding back upon the SEB itself. This feedback acts negatively on T_s in unstable conditions, since a small positive perturbation in T_s increases thermal instability and aerodynamic conductance, exerting a restoring influence on T_s through increased aerodynamic heat transfer.

3 *Physiological coupling* Vegetation physiology and the SEB exert strong short-term influences upon each other through T_s. This feedback acts positively upon T_s at temperatures high enough for an increase in T_s to cause a decrease in stomatal conductance. This can lead to complete stomatal closure (Monteith 1975; Jarvis 1976; Jarvis & McNaughton 1986; Collatz *et al.* 1991), but such a tendency is substantially mitigated in most circumstances by aerodynamic feedback (Raupach 1997). Root depth and vertical distribution is another important physical attribute of plants which influences soil water storage (Nepstad *et al.* 1994), that in turn affects the SEB through the partitioning of sensible and latent heat fluxes.

4 *CBL coupling* The saturation deficit (the difference between saturation and ambient humidities) in the air above the surface is not independent of the SEB during the day, because it evolves in response to the growth of the atmospheric CBL. This growth in turn is determined by the SEB (together with the temperature and humidity structure of the overlying troposphere), leading to a coupling mechanism which constitutes a powerful feedback control on the daytime behaviour of both the SEB and the CBL (McNaughton & Spriggs 1986, 1989; Raupach 1991; McNaughton & Raupach 1996). CBL coupling is generally a negative feedback process, since a perturbation of the SEB toward (say) increased sensible heat flux will increase the rate of entrainment of dry air from above and thence the saturation deficit in the CBL, in

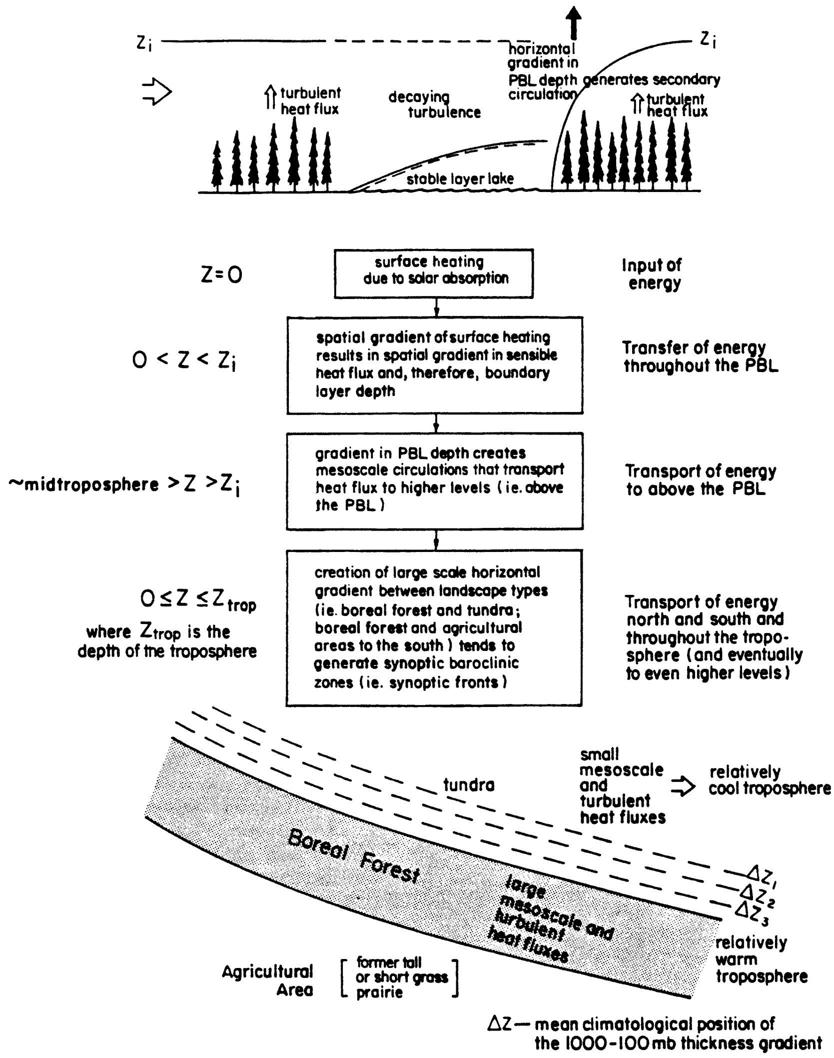

Fig. 1. Conceptual model of how the surface heterogeneities influence the circulations above them through differential fluxes and their associated boundary layers (from Pielke & Vidale 1995).

turn increasing the latent heat flux and decreasing the sensible heat flux at the surface. A second process important in CBL coupling is boundary-layer cloudiness, which constrains CBL growth by decreasing the available energy at the surface.

Effect of surface inhomogeneity

The Earth's land is composed of variable terrain elevation and landscape types. As shown in Atkinson (1981), Pielke (1984), and Cotton & Pielke (1995), for example, these spatial variations in land surface properties cause horizontal variations in the surface energy budget. These variations can generate mesoscale atmospheric circulations which can focus rainfall, and profoundly influence local and regional weather. Figure 1 illustrates how such surface inhomogeneity can feed upscale, even to the global scale.

Over horizontally homogeneous surfaces, the Monin–Obukhov similarity theory has been widely used for the estimation of the surface turbulent fluxes, and various theories have also been developed for the whole boundary layer (e.g. Stull 1988). When the spatial scale (L_s) of the surface inhomogeneities is small (e.g. less than 5 km), their effect on the turbulent fluxes is usually limited to the lower part of the boundary layer, and this effect can be considered through a modification to the Monin–Obukhov similarity theory (Avissar & Schmidt 1998). When L_s is sufficiently large (e.g. greater than 10 km), however, horizontal differential heating can lead to the development of mesoscale circulations which will affect the whole boundary layer.

These mesoscale circulations are not considered in the above methods. Furthermore, it is even unclear at present whether the surface fluxes in the cases of large L_s can be computed by the above methods.

The impact of landscape heterogeneity on atmospheric and surface (energy, water, and mass) fluxes, *F*, at local scales and mesoscales can be, in general, determined from the planetary boundary layer depth, the characteristic horizontal patch size, the potential temperature difference between boundary layer top and the surface, the large-scale wind, the surface potential temperature, the maximum horizontal surface temperature difference between different patches, the height above the ground, and the surface sensible heat, moisture, or momentum flux (Zeng & Pielke 1995a; Lynn *et al.* 1995). For small patches and strong large-scale winds, for example, the influence of landscape variability quickly homogenizes a short distance above the surface (Dalu *et al.* 1996).

Small spatial-scale surface inhomogeneity can still affect the detailed structure of atmospheric boundary layer, however (e.g. Hadfield *et al.* 1992; Walko *et al.* 1992; Zeng & Pielke 1993; Avissar & Schmidt 1998), but it can hardly generate mesoscale circulations, and the aggregated effect might be represented by the mosaic approach (e.g. Avissar & Pielke 1989; Koster & Suarez 1992; Giorgi 1997a,b) or by the use of blending heights with effective surface parameters (e.g. Mason 1988; Blyth *et al.* 1993).

At mesoscales, however, surface inhomogeneity can generate mesoscale circulations in the boundary layer. In other words, surface inhomogeneity can affect both surface fluxes and fluxes in (and even above) the boundary layer (e.g. Pielke *et al.* 1991; Avissar & Chen 1993; Chen & Avissar 1994; Lynn *et al.* 1995; Zeng & Pielke 1995a; Avissar & Liu 1996). It is demonstrated in Zeng & Pielke (1995b) that these mesoscale fluxes are insensitive to small perturbations in initial and surface boundary conditions as well as model parameters. In other words, they are parameterizable. Based on dimensionless parameters, parameterization schemes have been developed for turbulent and mesoscale fluxes in Zeng & Pielke (1995a). A complimentary approach has also been reported in Lynn *et al.* (1995) for the development of these parameterization schemes.

Mesoscale land surface variability and rainfall

Several studies have shown that mesoscale land surface variability can influence both the amount of precipitation and its spatial distribution (Andre *et al.* 1989b; Blyth *et al.* 1994; Chen & Avissar 1994; Avissar & Liu 1996). For stratiform precipitation, mechanisms causing an increase in rainfall include: frictional effects, which can cause a change in horizontal convergence and associated vertical velocities; and rapid evaporation of rainfall from the canopy of vegetation or bare soil, which may cause an increase in mixing ratio of the boundary layer. The first process also plays a role in the orographic enhancement of rainfall (e.g. Cotton & Anthes 1989). It is also possible that periodic changes in the land surface roughness may generate vertical motion, which in turn may trigger cumulus convection (Dalu *et al.* 1996). Such a process may well occur when, for instance, deforestation takes place in regular patterns, such as in the Amazon basin in the state of Rondonia.

Anthes (1984) suggested that the spacing of vegetation in semiarid regions can optimize the amount of cumulus convective precipitation that falls. This enhancement of precipitation occurs because of the development of local wind circulations that concentrate water vapour from transpiration in deep cumulus clouds.

The study by Blyth *et al.* (1994) attempted to separate the effects of moisture convergence and the rapid humidity feedback through a series of sensitivity experiments with a 3D mesoscale model. Their simulations referred to an area of 400×400 km in south-west France with both agricultural land and forest. They concluded that complete forest cover in the domain could increase the frontal rainfall by 30% compared to a simulation in which the domain was bare soil. About half of that increase was caused by moisture convergence, the other half was by rapid re-evaporation of intercepted rainfall. Especially in coastal areas, or areas with a large number of frontal intrusions, the frictional effects eventually represent a true gain in actual soil moisture content.

Avissar & Liu (1996) have shown how the specific pattern of land surface may affect the formation of shallow convective clouds and precipitation. Their study illustrates that thermal inhomogeneities in the land surface may trigger thermal circulations, which may bring moist air to high elevations where it can be precipitated, especially in semiarid regions, where surface energy partitioning is strongly dependent on antecedent rainfall. This process may lead to preferred and seasonally persistent precipitation patterns, while at the same time over a sufficiently large area, the total precipitation would not change significantly (Taylor *et al.* 1997). They suggested that locally increased potential energy in the boundary layer enhanced the convective precipitation at the scale of a single convective cell (10 km). This is a potential land surface atmosphere feedback at a much smaller scale than previously investigated with (hydrostatic) numerical models.

Continental scale processes: the Amazon example

The Amazon basin, due to its size and equatorial location, is a major heat and moisture source for the general circulation of the global atmosphere. The direct consequences of the Amazon deforestation include the change of vegetation type and soil properties, the increase of surface albedo, and the decrease of the roughness length. These changes, in turn, will alter the regional transfers of water, energy, and momentum between the surface and the atmosphere, which will possibly result in the increase of surface temperature and the decrease of precipitation and evaporation over the Amazon basin, as suggested by general circulation model (GCM) studies (e.g. Henderson-Sellers *et al.* 1993). Biomass burning and changes in the photosynthetic function of the vegetation will also add carbon dioxide to the atmosphere, which will enhance the global greenhouse effect (e.g. Tans *et al.* 1990). Furthermore, aerosol from biomass burning will affect the global radiation budget (Penner *et al.* 1992). Because of these effects and the biodiversity loss resulting from the deforestation, the Amazon basin has become one of the foci of interdisciplinary research in the areas of atmospheric science, hydrology, ecology, and biogeochemistry, as demonstrated by field experiments in the past and planned experiments in the near future (e.g. the Large Scale Biosphere–Atmosphere Experiment in Amazonia, LBA, for 1997–2000).

As mentioned above, most of the previous GCM studies showed that deforestation would reduce precipitation and evaporation, and increase surface temperature over the Amazon basin. These studies also indicated that deforestation could disturb the climate in the nearby regions, and likely disturb some aspects of the general circulation, especially the Walker and Hadley cells. In contrast, using the monthly mean outgoing longwave radiation (OLR) data from the NOAA polar-orbiting satellites, Chu *et al.* (1994) found little indication for a rainfall decrease associated with deforestation over the Amazon basin in the past 15 years. One possible reason for the different results between this observational study and the previous GCM studies is the assumption in GCM studies of deforestation over the whole basin which has not occurred yet in reality although large areas of Amazonia have been changed from forest to pasture and agricultural land (e.g. Skole & Tucker 1993). Furthermore, the study of Avissar & Liu (1996) seems to indicate that deforestation at the mesoscale, as is currently occurring in Amazonia, increases the recycling of water and does not reduce precipitation.

Three features related to the land surface have been identified as important to deforestation studies based on theoretical grounds and GCM results. Sellers 1992, (fig. 14.3) provides a list of possible feedbacks due to change in the three factors: albedo, evaporation (soil moisture in that figure), and surface roughness length. Although there are quite a few feedbacks in general, these are only a few important interactions with respect to Amazon deforestation (Zeng *et al.* 1996).

In order to facilitate the theoretical understanding of the regional and possible global impact of Amazon deforestation, an intermediate-level model is developed in Zeng *et al.* (1996) for tropical climatology including atmosphere–land–ocean interaction. Analysis of the thermodynamic equation in the model reveals that the balance

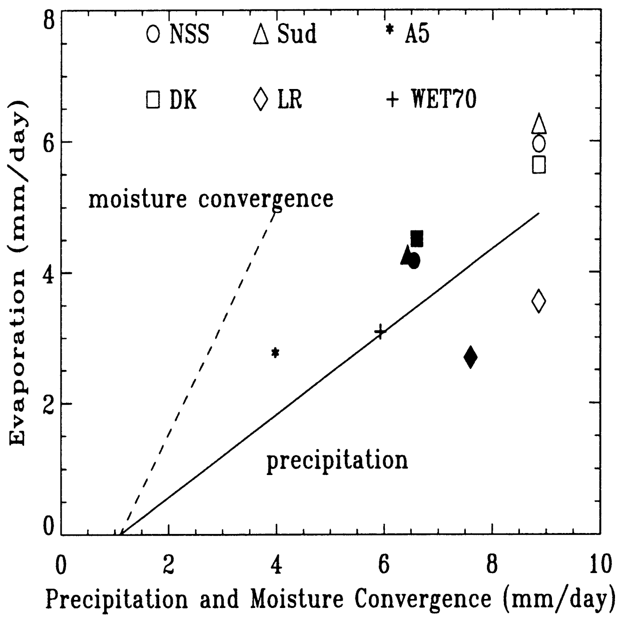

Fig. 2. Relationship of precipitation and moisture convergence with evaporation for the continuous deforestation experiment denoted by the solid and dashed lines, respectively. Also plotted are some GCM deforestation simulation results of evaporation and precipitation scaled by the control precipitation of the intermediate-level model. NSS, Nobre *et al.* (1991); DK, Dickinson & Kennedy (1992); LR, Lean & Rowntree (1993); Sud, Sud *et al.* (1996). A5 and WET70 are two deforestation experiments (by increasing albedo by 0.05 and by decreasing surface wetness by 30%, respectively) using the intermediate-level model. Open symbols are for the control runs and filled symbols are for deforestation (from Zeng *et al.* 1996).

between convective heating, adiabatic cooling, and radiation largely determines the deforestation response. Figure 2 shows the quasi-linear relationship between precipitation, moisture convergence, and evaporation under a continuous deforestation scenario. Also plotted in the figure are Amazon deforestation results from some of the previous GCM experiments. The control runs of these GCM experiments have large variations, so the precipitation and evaporation are scaled so that the control precipitations are all the same as that of the model of Zeng *et al.* (1996). With such a scaling, three of the four GCM results are very close to each other. Without it, differences between their control runs are often larger than the differences between control and deforestation runs. Even though the results in Lean & Rowntree (1993) are somewhat different from those in the other three studies, the slope linking the open and filled symbols in Fig. 2 for each GCM study is similar among the four studies and is similar to that of the intermediate-level model of Zeng *et al.* (1996). The implication is that the highly nonlinear land–atmosphere (including the recycling of precipitation) interaction and large-scale dynamics work together to maintain a quasi-linear relation between precipitation, moisture convergence, and evaporation.

Sensitivity studies (atmospheric response to changes in boundary forcing)

The sensitivity of global-scale atmospheric circulation and surface climate to changes in land-surface boundary conditions has been consistently demonstrated for more than 20 years by performing experiments with atmospheric general circulation models. Traditionally, a simulation is performed after changing some aspect of the land surface (albedo, roughness, evapotranspiration, biome distribution), and the results are compared with a control simulation. This approach was used first by Charney (1975), who showed that the dynamical response of atmospheric circulation to changes in surface albedo acted as a positive feedback mechanism to enhance and sustain drought conditions in sub-Saharan Africa. Using this forcing-response approach, the sensitivity of atmospheric circulation and climate has been investigated in terms of changes in terrestrial albedo (Chervin 1979; Sud & Fennessy 1982), evapotranspiration (Shukla & Mintz 1982) surface roughness (Sud & Smith 1985; Sud *et al.* 1988), tropical vegetation cover (Dickinson & Henderson-Sellers 1988; Nobre *et al.* 1991; McGuffie *et al.* 1995), soil moisture (Delworth & Manabe 1989), boreal forest cover (Bonan *et al.* 1992), stomatal conductance (Henderson-Sellers *et al.* 1995; Pollard & Thompson 1995), and leaf area index (Chase *et al.* 1996).

This approach clearly demonstrated the influence of terrestrial vegetation on weather and climate, but such sensitivity experiments do not allow for two-way interactions between ecosystems and the atmosphere: arbitrary changes were imposed in the form of static boundary conditions. Rather, they demonstrate the response of one component (the atmosphere) to prescribed changes in the forcing related to another component (the land surface) of the system. Moreover, in most cases changes to a particular surface parameter (such as roughness length or albedo) were made without regard for biophysical realism: such changes in nature do not occur without concurrent changes in other parameters (such as leaf area index, stomatal conductance, etc.). In the 1980s, new parameterizations of the interactions between the land surface and the atmosphere were constructed to retain a greater degree of biophysical consistency among the physical properties of the vegetation (Dickinson 1984; Dickinson *et al.* 1986; Sato *et al.* 1989; Sellers *et al.* 1986). In these models, biophysical properties were still prescribed noninteractively, but an attempt was made to maintain internal consistency among radiative, aerodynamic, and hydrological properties of the vegetation by using physiologically based representations of the vegetated land surface.

A new approach to the study of short-term (seconds to weeks) and medium-term (months to years) interactions between the atmosphere and terrestrial ecosystems has recently been adopted by some investigators. A key parameter in the calculation of latent and sensible heat flux, and its integration from the scale of individual leaves to plant canopies to model grid cells, is the canopy integrated stomatal conductance to water vapour. A growing number of modelling groups (Bonan *et al.* 1995; Sellers *et al.* 1996a,b,c; Randall *et al.* 1996; Berry *et al.* 1997; Foley *et al.* 1997) are now calculating stomatal conductance by introducing the photosynthetic assimilation of atmospheric CO_2 and the biochemical and physiological feedbacks that relate photosynthesis, stomatal conductance, and transpiration (Farquhar 1979; Farquhar *et al.* 1980; Ball 1988; Collatz *et al.* 1991, 1992). This approach is based on an empirical relationship between stomatal conductance and the rate of carbon fixation by photosynthesis which appears to be useful for a wide range of species, which recognizes the physiological optimization problem of minimizing water loss and maximizing carbon gain. Leaf-level stomatal conductance is integrated to the canopy scale based on nitrogen allocation (Cowan 1986; Field & Mooney 1986; Givnish 1986), which follows the long-term distribution of light in the canopy (Sellers *et al.* 1992b) and so can be deduced from remotely sensed data.

A distinct advantage of the new generation of land-surface models is their ability to capture some of the effects of nonlinear interactions between climate and terrestrial ecosystems. Sellers *et al.* (1996c) used such a model in an atmospheric GCM to study the equilibrium response of the biosphere–atmosphere system to a doubling of atmospheric CO_2. In addition to the radiative effects of elevated CO_2 in the GCM, their results showed a shift in the local surface energy budget toward sensible heating because plants could assimilate more carbon for a given stomatal conductance, so evapotranspiration was reduced. More generally, the atmospheric and biological responses were characterized by nonlinear interactions which could not be predicted in a simpler forcing-response model.

Another important interaction between terrestrial ecosystems and atmospheric circulation has been explored by Denning *et al.* (1995, 1996a,b), who showed that nonlinear interactions between photosynthesis and boundary-layer turbulence may have a first-order effect on the distribution of atmospheric CO_2. Nonlinear interactions between the atmosphere and terrestrial biosphere also occur on local to regional scales due to the covariance between atmospheric convection and photosynthesis (Denning *et al.* 1995, 1996a,b). Both buoyancy-driven turbulence and convection in the atmosphere and photosynthetic carbon assimilation in the biosphere are driven by solar radiation. This means that on average, photosynthesis acts on a deeper layer of the atmosphere than respiration, because it occurs during the day and during the growing season when radiation also produces a positive surface energy balance and a deep ABL. Conversely, respiration dominates at night and during the colder months of the year when the ABL is shallow and stable. This covariance produces a vertical gradient in atmospheric CO_2 over the vegetated land areas of the world in the annual mean by systematically venting low-CO_2 air during times of high photosynthesis and retaining respiration-influenced air near the ground under stable inversions. This 'rectifier' effect may lead to a meridional gradient in the zonal mean CO_2 near the surface which is half as strong as that produced by fossil fuel emissions (Denning *et al.* 1995). The apparent meridional gradient is produced because the biological signal is rectified over seasonal vegetation, which is predominantly found in the northern hemisphere, and because the observational data on atmospheric CO_2 are mostly confined to near-surface observations.

Conclusion

From existing modelling and observational studies, the evidence is convincing that short-term biophysical and long-term biogeochemical effects significantly influence weather and climate. Terrestrial ecosystem dynamics, are therefore, an influential component of the Earth's climate system and in studies of past and possible future climate change, its' variations are as important as changes in atmospheric dynamics and composition, ocean circulation, ice sheet extent, and orbital perturbations.

Acknowledgements

This paper is a product of a workshop on 'Bidirectional Ecosystem–Atmosphere Interactions' held in San Jose dos Campos, Brazil on 8–11 July 1996. The meeting was part of the core program of FOCUS 3 of the International Geosphere-Biosphere Program (IGBP) Core Project on Biospheric Aspects of the Hydrological Cycle (BAHC), and of the International Satellite Land Surface Climatology Program (ISLSCP). It was jointly funded by NSF under grant ATM-9634021, NASA under grant NAGW-5226, EPRI, and the IGBP Biospheric Aspect of the Hydrological Cycle (BAHC) Core Project Office. R. Pielke's work reported here is supported by the NSF under grant ATM-9306754 and the USGS under contract 1431 94-A-01275. R. Avissar's work is supported by NSF under grant ATM EAR9415441 and NOAA under grant NA56GP0200. Xubin Zeng's work was supported by NSF under grant ATM-9419715. A.S. Denning is supported by NASA contract NAS5–31730.

We wish to thank the anonymous Referees and the Subject Editor for their excellent comments and suggestions which we adopted in our revised paper.

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