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REVIEW

THE EVOLUTION OF, AND REVOLUTION IN, LAND SURFACE SCHEMES DESIGNED FOR CLIMATE MODELS

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ABSTRACT

The land surface is a key component of climate models. It controls the partitioning of available energy at the surface between sensible and latent heat, and it controls the partitioning of available water between evaporation and runoff. The land surface is also the location of the terrestrial carbon sink. Evidence is increasing that the influence of the land surface is significant on climate and that changes in the land surface can influence regional- to global-scale climate on time scales from days to millennia. Further, there is now a suggestion that the terrestrial carbon sink may decrease as global temperatures increase as a consequence of rising CO_2 levels. This paper provides the theoretical background that explains why the land surface should play a central role in climate. It also provides evidence, sourced from climate model experiments, that the land surface is of central importance. This paper then reviews the development of land surface models designed for climate models from the early, very simple models through to recent efforts, which include a coupling of biophysical processes to represent carbon exchange. It is pointed out that significant problems remain to be addressed, including the difficulties in parameterizing hydrological processes, root processes, sub-grid-scale heterogeneity and biogeochemical cycles. It is argued that continued development of land surface models requires more multidisciplinary efforts by scientists with a wide range of skills. However, it is also argued that the framework is now in place within the international community to build and maintain the latest generation of land surface models. Further, there should be considerable optimism that consolidating the recent rapid advances in land surface modelling will enhance our capability to simulate the impacts of land-cover change and the impacts of increasing CO_2 on the global and regional environment. Copyright © 2003 Royal Meteorological Society.

KEY WORDS: land surface modelling; climate modelling; review; global scale; carbon

1. INTRODUCTION

A major review paper recently published in this journal provides a detailed and comprehensive discussion of climate modelling (McGuffie and Henderson-Sellers, 2001). Climate models are used in applications ranging from experiments into stratospheric processes through to investigation of the deep ocean circulation. One of the most common uses of climate models is to explore the impact of perturbations caused by human activity. Increasing carbon dioxide (CO_2) is one such example, and the Intergovernmental Panel on Climate Change (IPCC; Houghton *et al*., 2001) provide a comprehensive assessment of this area, including an assessment of the performance of climate models (McAvaney *et al*., 2001). Humans have also altered a significant fraction of the Earth's surface via land-cover change (LCC) (Vitousek *et al*., 1997). Human-induced LCC is likely to accelerate in the 21st century as direct impacts via reafforestation, deforestation or agricultural intensification become supplemented by indirect effects of human activity. This indirect effect, via CO_2 fertilization or

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CO_2-related climate change, is expected to lead to global-scale changes in the structure, density and function of plants.

Within a climate model, the element that simulates the initial effect of LCC is the land surface model (LSM). The LSM is a key component in understanding the Earth's carbon cycle and how CO_2 increases in the atmosphere may be moderated by natural terrestrial carbon sinks (Prentice *et al*., 2001). In order to project future climates, the way that the Earth's surface interacts with the atmosphere, and the ways that this interaction changes as a result of both human activity and natural processes, must be represented. Similarly, in order to simulate the impacts of deforestation, reafforestation or agricultural intensification using a climate model, the LSM must be able to represent the impact of these changes on surface–atmosphere interactions.

This paper, therefore, considers one element of climate models, the land surface, and describes why the land surface should play a key role in climate models. The terms of reference for this review are intentionally limited to land surface processes in *climate models*. Issues relating to micro- to meso-scale processes, weather forecasting, etc., will not be considered (these deserve full attention in their own right). The definition of the 'land surface' used here is the surface that comprises vegetation, soil and snow, coupled with the way these influence the exchange of water, energy and carbon within the Earth system. Following a theoretical examination of why the land surface should matter in climate and climate models (Section 2), evidence that it does play a major role will be presented (Section 3). This will be followed by a description and examination of the historical development of LSMs (Section 4). Finally, outstanding issues will be identified and the direction of LSM development over the next few years will be foreshadowed (Section 5).

2. WHAT DOES THE LAND SURFACE NEED TO REPRESENT IN CLIMATE MODELS?

There are several fundamental equations that represent the key role played by the surface in climate. The two key equations represent the surface energy balance and the surface water balance. The surface also strongly influences momentum exchange and, perhaps most critically for the future climate, the carbon balance.

2.1. The surface energy balance

The shortwave radiation emitted by the Sun is reflected, absorbed or transmitted by the atmosphere. An amount of energy $S\downarrow$ reaches the Earth's surface and some is reflected (depending on the albedo α). Of 100 units of energy entering the global climate system, 46 are absorbed by the surface (Figure 1, Rosen, 1999). Infrared radiation is also received ($L\downarrow$) and emitted ($L\uparrow$) by the Earth's surface (depending on the temperature and emissivity of the land and atmosphere). The net balance of the incoming and reflected shortwave radiation, and the incoming and emitted longwave radiation at the Earth's surface is called net radiation R_n:

$$R_n = S\downarrow(1-\alpha) + L\downarrow - L\uparrow \tag{1}$$

Of the 100 units of energy entering the global climate system, 31 are exchanged as sensible and latent heat fluxes (Figure 1, Rosen, 1999), otherwise known as the turbulent energy fluxes. The land surface significantly influences the way that these 31 units of energy are partitioned between sensible H and latent heat λE fluxes, and also acts as a significant medium to store energy on diurnal, seasonal and longer time scales (thousands of years in the case of heat stored in permafrost). R_n must be balanced by $H, \lambda E$, the soil heat flux G and the chemical energy F stored during photosynthesis and released by respiration (which is usually omitted in climate models as it amounts to less that 1% of absorbed insolation; Sellers, 1992):

$$R_n = H + \lambda E + G + F \tag{2}$$

Changes in the surface albedo affect R_n, and thus H and λE. Albedo changes naturally with solar insolation angle, seasonally with vegetation changes and stochastically with rain or snowfall. It can also be changed directly via human activity (by LCC for example) or indirectly (via the fertilization effect of increasing CO_2,

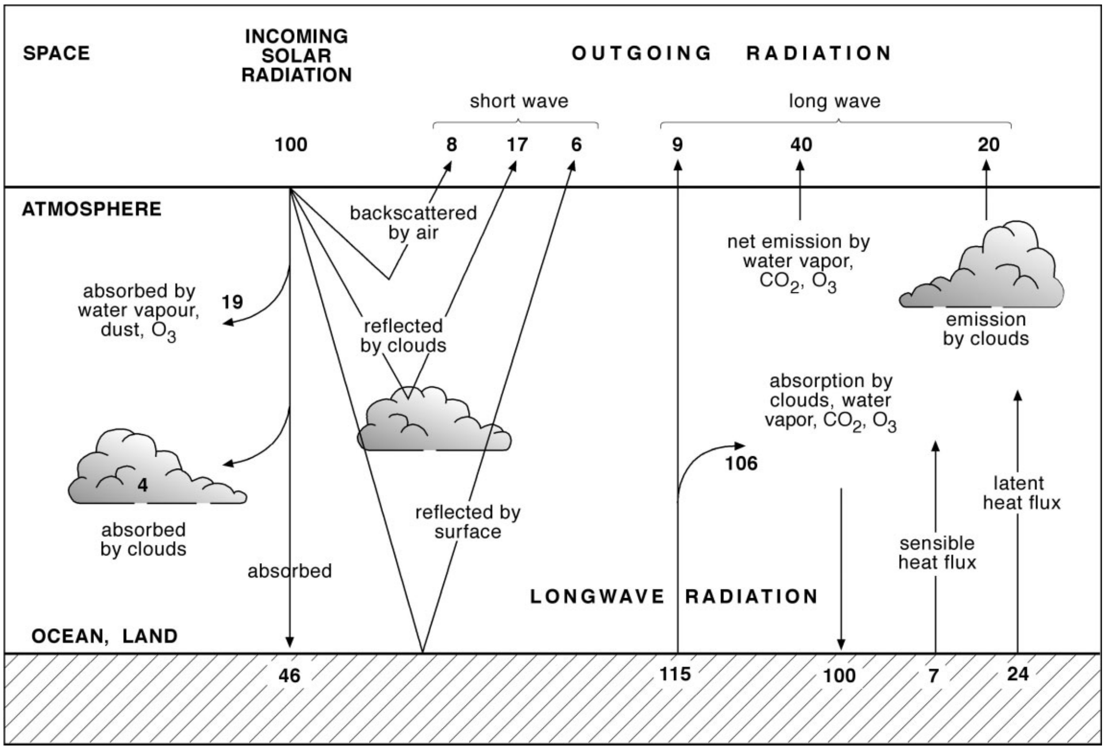

Figure 1. Schematic diagram of the annual mean global energy balance. Units are percentage of incoming solar radiation. The solar (shortwave) fluxes are shown on the left-hand side and the terrestrial (longwave) fluxes are on the right-hand side (redrawn from Rosen, 1999)

which may change the physical characteristics of vegetation). In terms of climate modelling, it is important to partition R_n between H and λE as well as possible, since less λE contributes less water vapour to the atmosphere and tends towards decreasing cloudiness and precipitation, whereas decreases in H tend to cool the planetary boundary layer and reduce convection (see Betts *et al*., 1996 and Figure 2). Complex feedbacks exist whereby changes in clouds or precipitation feedback to modify the initial perturbation to albedo (Figure 2). Given the key role that H and λE play in the climate system, it is necessary to simulate the diurnal, seasonal and longer term variations in these fluxes as well as possible. This has become a key focus in how land surface processes are represented in climate models.

H and λE are sensitive to the nature of the land surface in many ways (Verstraete and Dickinson, 1986). Changes in the actual vegetation cover alter the surface area of vegetation in contact with the atmosphere and the balance between fluxes from the soil and vegetation. Changes in the leaf area index (LAI, the surface area of leaf per surface area of ground) can influence the exchange of both H and λE (Figure 3). Changes in the distribution of roots can dramatically change the amount of soil moisture available to plants to transpire (Figure 4) and a positive feedback between reduced root water uptake, rainfall and further reductions in root depth may exist (Figure 4). Given that H and λE are turbulent energy fluxes, the aerodynamic roughness of the surface can exert a strong influence, and, again, a feedback exists between changes in precipitation, vegetation and roughness length (Figure 5).

The link between surface characteristics and the turbulence that drives the exchange of H and λE is the aerodynamic resistance r_a. This turbulent diffusion term also affects the transfer of CO_2 away from the vegetation surface to the free atmosphere. There is a different aerodynamic resistance term for momentum transfer (see Garratt, 1993), which is not discussed here. The aerodynamic resistance is inversely dependent upon the wind speed and the logarithm of the surface roughness length, which, in turn, is a function of the drag properties of the land surface. Stability corrections (see Garratt, 1993) need to be applied to account for

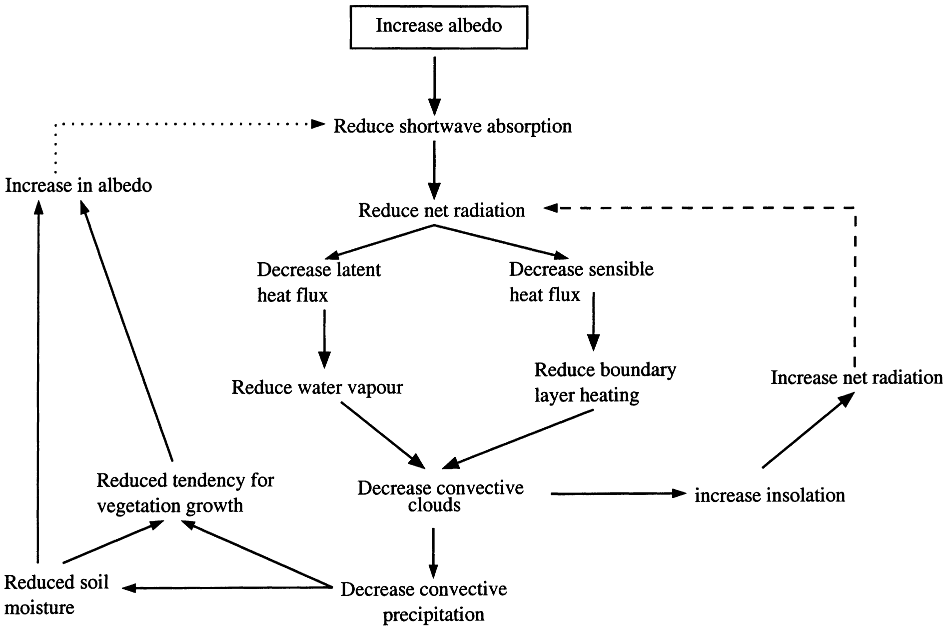

Figure 2. Conceptual diagram of the impact an increase in albedo has on the land surface and some elements of the boundary-layer climate. The dotted line represents a positive feedback and the dashed lines represent a negative feedback

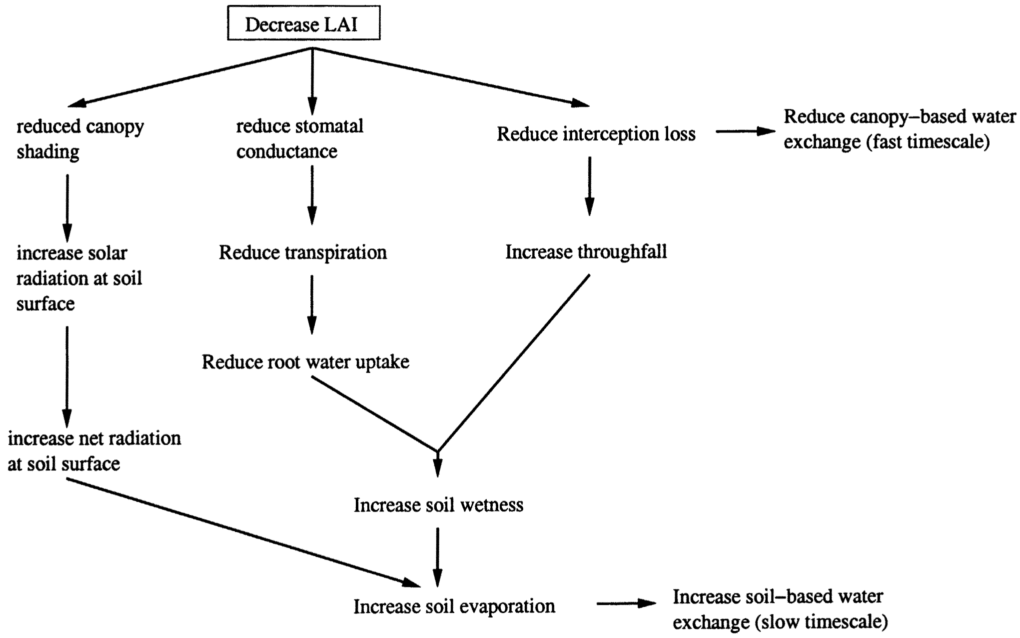

Figure 3. As Figure 2, but for the LAI

the effects of convection on r_a. Changes in the surface roughness length directly affect r_a (Figure 5). Since rough surfaces (e.g. forests) are more tightly coupled to the atmosphere than a smooth surface (e.g. grass), a higher roughness length permits a greater exchange of H and λE for a given set of meteorological conditions. If the nature of the surface changes, the change in surface roughness will directly affect the exchange of H,

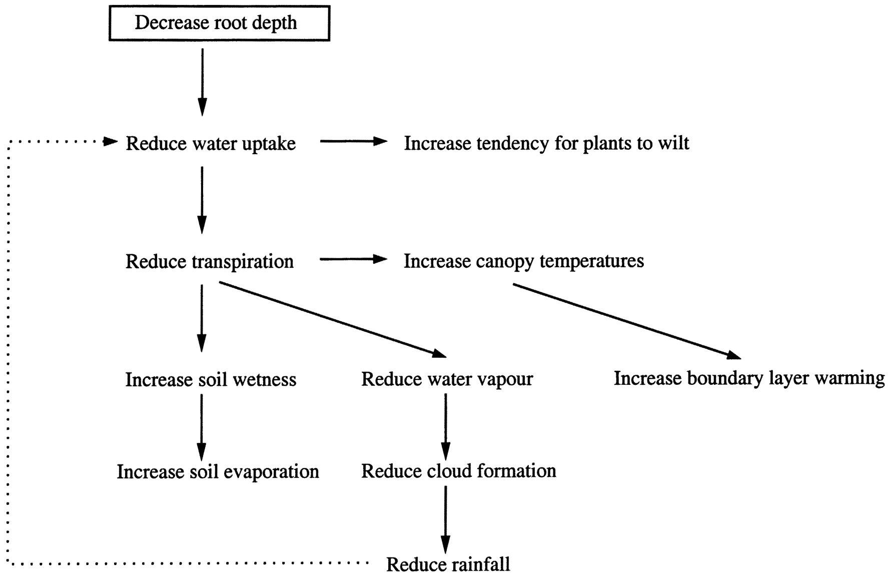

Figure 4. As Figure 2, but for roots

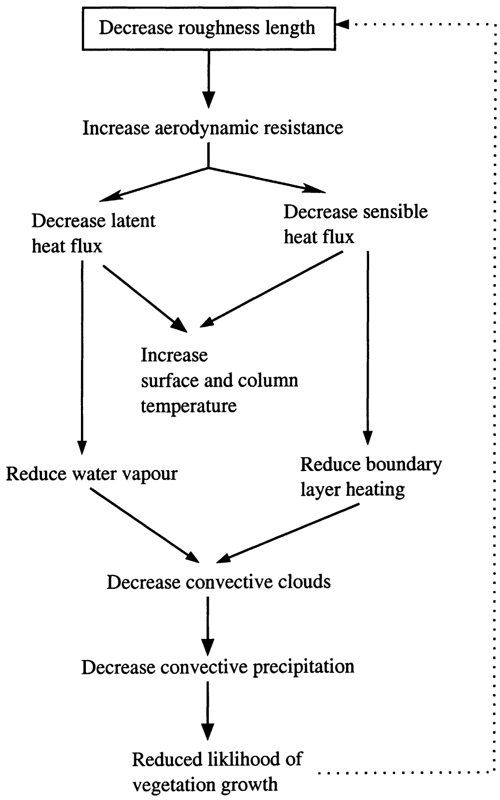

Figure 5. As Figure 2, but for the aerodynamic roughness length

λE, CO_2 and momentum. Mathematically, H can be represented as a quasi-diffusive process:

$$H = \frac{T_s - T_r}{r_a}\rho c_p \tag{3}$$

where T_s is the surface temperature (itself a function of the surface energy balance), T_r is a reference temperature above the surface, ρ is the air density and c_p is the specific heat of air. λE is a more complex process than H, as it involves most of the difficulties of H plus all those processes that enable the water to evaporate (canopy interception, root water uptake, soil moisture diffusion, etc.). Although there are several ways to represent λE in climate models, the aerodynamic approach is commonly used (Sellers, 1992):

$$\lambda E = \left(\frac{e^*(T_s) - e_r}{r_s + r_a}\right)\frac{\rho c_p}{\gamma} \tag{4}$$

where $e^*(T_s)$ is the saturated vapour pressure at T_s, e_r is the vapour pressure at a reference height and γ is the psychrometric constant. The remaining term, r_s, is the surface resistance to the transfer of water from the surface to the air. This resistance includes the resistance to water evaporating from the soil through stomates, or within the plant. The inclusion of stomates in the surface resistance means that λE from vegetated surfaces is tightly coupled to biological activity through photosynthetic activity, which, in turn, is linked to carbon uptake by plants.

The soil heat flux G is a diffusion–conduction process, which can be generalized as:

$$G = -K\,\mathrm{d}T_s/\mathrm{d}z \tag{5}$$

where z is soil depth and K (W m^{-1} K^{-1}) is thermal conductivity. This simplifies the complexity of soil heat conduction, which is usually treated with care in climate models following, for example, Hillel (1982).

Although climate models have traditionally ignored the photosynthesis term F, it will become clear later that this is no longer an appropriate simplification. To a reasonable degree of accuracy, the flux density (kg m^{-2} s^{-1}) of CO_2 is:

$$F = \frac{c_i - c_a}{r_{st} + r_a} \tag{6}$$

where c_i is the CO_2 concentration within the leaf and c_a is the ambient CO_2 concentration. This expression takes into account the resistance to the transfer of CO_2 from the atmosphere, through the stomates to the chloroplast, the stomatal resistance r_{st}. Note that r_{st} is not identical to, but is related to, r_s in Equation (4).

Thus, the key characteristics of the land surface, in terms of influencing the exchange of H, λE and CO_2 with the atmosphere, are the albedo, roughness length and the characteristics of plants that influence their surface area (LAI) or their ability to take up water from the soil (roots) and transpire it (LAI, and the resistance exerted by stomates on transpiration and CO_2 uptake). A major impact of changes in the nature of the land surface is the effect on the time scale of surface–atmospheric exchanges. Extremes, especially temperature, are affected by the nature of the surface and whether moisture can be supplied for evaporation and cooling. Thus, modifications in the nature of the land surface could be expected to affect not only mean surface–atmospheric exchanges, but also the extremes and the time scale of the response of the land surface to various external perturbations.

2.2. *The surface water balance*

Precipitation that falls to the Earth's surface is either intercepted by vegetation or reaches the soil surface directly. Precipitation that is intercepted either evaporates or drips to the surface, and the drip, combined with the rainfall that reaches the surface directly, either infiltrates or runs across the soil surface (surface runoff). Water that infiltrates may evaporate from the soil surface, drain through the soil, or be taken up by roots and transpired. While the water is in the soil it is known as soil moisture, and although only a tiny proportion of the total water is stored in the soil (0.0012%, Oki, 1999) it plays a vital role in the provision of food and fresh water.

A basic role of the land surface is to partition available water (assumed here to be precipitation, P, but it could also include snow melt) between evaporation E and runoff, which is usually split between a fast component (R_{surf}) and a slow component (R_{drain}). Overall, this balancing of incoming and outgoing fluxes of water is called the surface water balance:

$$P = E - R_{\mathrm{drain}} - R_{\mathrm{surf}} - \Delta S \tag{7}$$

where ΔS is the change in soil moisture storage. Note that the hydrological cycle uses E (kg m^{-2} s^{-1}); this is linked to the surface energy balance, which uses λE (W m^{-2}), where λ (J kg^{-1}) is the latent heat of vaporization. The two runoff terms are influenced by the characteristics and wetness of the soil. There are a variety of ways of modelling these processes, and these will be discussed later.

Changes in the characteristics of the land surface affect the surface water balance. In particular, a change in the nature of vegetation affects interception and transpiration (Figures 2–5). A change in the distribution of vegetation modifies the balance between fluxes originating from the soil and those derived through canopy processes. Changes in evapotranspiration, soil evaporation, re-evaporation of intercepted water, etc. affect runoff and soil moisture content. These then affect a variety of other processes through the link with the surface energy balance (Figure 6).

2.3. *The climatic effect of snow*

Over 50% of Eurasia and North America can be seasonally covered by snow (Robinson *et al*., 1993), leading to significant spatial and temporal fluctuations in surface conditions. The properties of snow (e.g. high albedo, low roughness length and low thermal conductivity) lead to impacts at the global scale (Vernekar *et al*., 1995). Foster *et al*. (1982) showed that, over the Eurasian continent, between 18 and 52% of variance in winter temperature could be explained by the extent of snow cover present in autumn, and that extensive snow cover over continental areas leads to the development of anticyclonic conditions (see also Barnett *et al*., 1989). In effect, snow has a substantial impact on climate (Cohen and Rind, 1991).

Snow is also one of the key feedbacks within the climate system and plays a very strong role as a positive feedback, enhancing the initial impacts of perturbations on the land surface (Cess *et al*., 1991; Randall *et al*.,

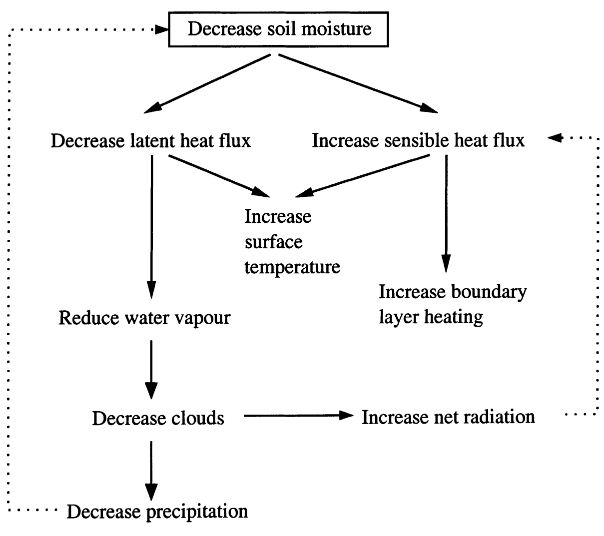

Figure 6. As Figure 2, but for soil moisture

1994; Betts, 2000). The snow–albedo feedback, whether coupled with changes in vegetation (Betts, 2000) or not (Colman *et al*., 1994), is a strong amplification to warming caused by increasing CO_2. The representation of snow in climate models has, therefore, been seen as a priority since the first climate models were constructed.

2.4. Carbon

The Earth's land surface stores very large amounts of carbon, but estimates vary considerably. For example, Pielke *et al*. (1998) estimate 1600 Gt of carbon and Prentice *et al*. (2001) estimate 2000 Gt of carbon (1 Gt = 1 Pg = 10^9 kg of carbon). These estimates are more than twice the store of carbon in the atmosphere as CO_2. The annual exchange of carbon between the terrestrial and atmospheric systems varies remarkably. In the 1980s the carbon exchange was about -0.2 ± 0.7 Gt year^{-1} (a net uptake; Prentice *et al*., 2001), but from 1990 to 1999 the net land–atmosphere carbon exchange was -1.4 ± 0.7 Gt year^{-1}. This net uptake of carbon by the terrestrial biosphere includes a natural sink due to biological activity and a human-enhanced source due to land clearance. During the 1980s, the sink into the biosphere was between -3.8 and $+0.3$ Gt year^{-1} (believed to be caused by some types of LCC, mainly Northern Hemisphere reafforestation, nitrogen fertilization and the fertilization effect of increased CO_2). This was counteracted by a carbon release of approximately 0.6–2.5 Gt year^{-1} due to other types of LCC (mainly deforestation in the tropics) (Prentice *et al*., 2001). What is clear is that a vast amount of carbon is exchanged between the Earth's surface and the atmosphere naturally, and that human activity is now modifying this exchange directly via LCC and indirectly through the fertilization effect of increased CO_2.

A major purpose of climate models is to simulate the evolution of climate over the next century or two. Given that CO_2 concentrations will increase over that time period, and that the increase is a major forcing mechanism on climate (Houghton *et al*., 2001), the ability to represent the impact of land surface processes on atmospheric CO_2 concentrations, and possible long-term net sources or sinks resulting from changes in the biosphere, are priorities in LSMs. A thorough and contemporary review of the carbon cycle is provided by McGuire *et al*. (2001) and Prentice *et al*. (2001).

2.5. Summary

Thus, there are good theoretical reasons why the land surface should affect the climate and the simulations by climate models via changes in H, λE, temperature, runoff, carbon and momentum transfer. Section 3 documents evidence to support this theoretical assessment.

3. EVIDENCE THAT THE LAND SURFACE DOES MATTER IN CLIMATE MODELS

3.1. Introduction

Several recent review papers have addressed the issue of whether the state of the land surface can influence weather and climate (Avissar and Verstraete, 1990; Betts *et al*., 1996; Pielke *et al*., 1998). There is now strong evidence that the land surface influences weather and climate at a range of time scales, from seconds to millions of years (Sellers, 1992; Pielke *et al*., 1998). The role played by the land surface within *climate* and *climate models* is different from the role played by the land surface in microclimatology, mesoscale meteorology, weather, etc. There are, of course, strong links across these scales, but, as far as we know, *climate* as modelled by *climate models* is not affected by small-spatial-scale processes operating over short time scales (e.g. Avissar and Pielke, 1991). Though there is good evidence from small time and space scales that land surface processes affect the atmosphere (Pielke, 2001), this evidence does not demonstrate that the integrated effects of these small-scale processes affecting the *atmosphere* locally, on time scales of a few days, actually affect *climate*. Further, evidence from observational studies (e.g. André *et al*., 1988; Sellers *et al*., 1992a, 1995; Goutorbe *et al*., 1997; Lyons, 2002), in themselves, does not provide proof that land surface processes matter to climate, as the evidence is limited in scale both spatially and temporally. What we need in order to establish the importance of the land surface is evidence from longer time scales and larger spatial scales.

3.2. Evidence that the land surface does matter in climate models

At regional to global scales, changes in the land surface influence climate. Evidence to support this statement comes from sensitivity studies that explore the impact of modifying the characteristics of the land surface. Mintz (1984) reviewed early experiments and provided the first overview that the land surface mattered to a relatively skeptical climate modelling audience. In terms of albedo, some of the key work demonstrating the sensitivity of climate to land surface albedo was performed by Charney *et al*. (1977), Cunnington and Rowntree (1986), Laval (1986), Sud and Fennessy (1982), Lofgren (1995) and, most recently, coupled with changes in greenhouse forcing, by Betts (2000). The significance of roughness has also been well researched (Sud and Smith, 1985; Sud *et al*., 1988). More recent studies (Chase *et al*., 1996) have focused on the role of LAI, finding a change in global temperature and precipitation following a prescribed change. The sensitivity of the climate (and changes in climate) to the water-holding capacity of the soil have been investigated by Milly and Dunne (1994), Milly (1997) and Ducharne and Laval (2000). A new area of recognition in climate models is the role of roots. De Rosnay and Polcher (1998) explored modelling root water uptake in a complex LSM coupled to a climate model and found some climate sensitivity to how roots were handled. Kleidon and Heimann (1998, 2000), and Zeng *et al*. (1998) also found sensitivity to changes in root depth. Feddes *et al*. (2001) review the role of roots in hydrological and climate models in more detail.

These experiments explored the sensitivity of the climate to a single parameter change. There is considerable additional evidence of the significance of land surface processes generated through regional-scale perturbation experiments. Work on deforestation (e.g. Bonan *et al*., 1992; Henderson-Sellers *et al*., 1993; Polcher and Laval, 1994; Lean and Rowntree, 1997), desertification (e.g. Xue, 1997; Nicholson *et al*., 1998) and land-use change (e.g. Chase *et al*., 1996, 2000; Betts, 2000; Zhao *et al*., 2001a,b) all point to large and statistically significant continental-scale changes, in temperature, rainfall and other variables as a result of LCC. Some of this work (Chase *et al*., 1996, 2000; Gedney and Valdes, 2000; Zhao *et al*., 2001a,b) provides evidence of large-scale atmospheric adjustment caused by regional-scale perturbations, leading to geographically remote changes, in temperature and precipitation. Very recent work suggests that LCC may affect extremes in temperature (Collatz *et al*., 2000) and precipitation (Zhao and Pitman, 2002). At long time scales, Claussen *et al*. (1998, 2001) have explored the role of vegetation–climate interactions. There has also been some work to explore the sensitivity of climate models to hydrology (e.g. Ducharne *et al*., 1998) and to the surface energy balance (Desborough *et al*., 2001).

The feedback between atmospheric and land surface processes over seasonal time scales has been discussed by Eastman *et al*. (2001) and Lu *et al*. (2001). Eastman *et al*. (2001) showed that the biological effect of doubling CO_2 on the grasslands of the central USA would be significant on the regional climate. Evidence also exists that representing the physiological and structural feedback of increasing CO_2 on the biosphere is important (Henderson-Sellers *et al*., 1995; Pollard and Thompson, 1995; Sellers *et al*., 1996; Betts *et al*., 1997; Levis *et al*., 2000; Cox *et al*., 2000; Friedlingstein *et al*., 2001). The two major feedbacks of increasing CO_2 on vegetation are structural (increase in LAI, roots, etc.) and physiological (reduced stomatal conductance, increased stomatal resistance). Figure 7 shows some of the complexity in these feedbacks. The physiological effect tends to reduce transpiration (but not reduce the uptake of CO_2), leading to a positive feedback on the warming caused by the increase in CO_2. The structural feedback tends to cool via an increase in evaporation and, combined with the increase in CO_2 uptake, acts as a negative feedback on warming induced by an increase in CO_2. It is not yet clear whether the net balance of these two effects is to warm or cool, given that our capability to represent these mechanisms in climate models is very new.

At global scales, climate models provide a significant amount of evidence that the land surface affects climate. Harvey (1988, 1989a,b), for example, showed that LCC during the last 7000 years amplified climate variations regionally and globally. Feedbacks, including the vegetation–snow–albedo feedback, appear to be important (Foley *et al*., 1994), since this positive feedback reduces surface albedo and, therefore, increases near-surface temperatures; this, in turn, favours growth of taller vegetation, thus reducing surface albedo

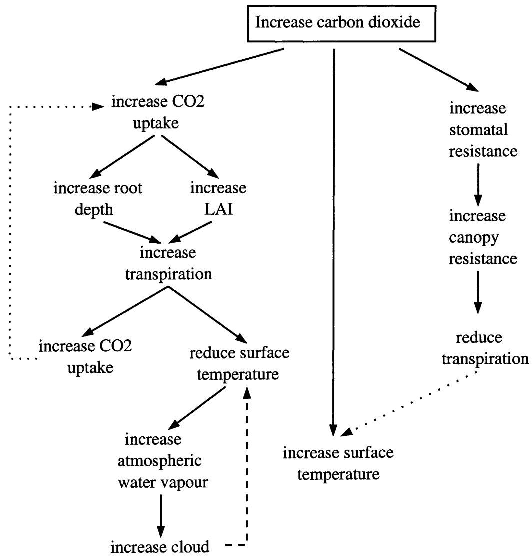

Figure 7. As Figure 2, but for the impact of increased CO_2 on a land surface model

further (see Figure 1; also see Claussen and Gayler, 1997 and Texier *et al.*, 1997). Other evidence, from Harvey (1989b), Berger *et al.* (1993) and de Noblet *et al.* (1996), shows that the vegetation–snow–albedo feedback contributed significantly to Northern Hemisphere cooling during the Eemian. Evidence that North Africa was much greener than today in the mid-Holocene can be explained, in part, via a positive feedback between vegetation and precipitation in this region (Kutzbach *et al.*, 1996; Texier *et al.*, 1997; Claussen and Gayler, 1997; Broström *et al.*, 1998). Overall, the role of the biosphere in explaining long-term (millennia-scale) climate changes is increasingly well established.

3.3. Summary

The evidence is now very strong that regional-scale land surface perturbations cause continental-scale changes in climate. There is also evidence that regional-scale land surface changes in key regions can cause significant changes in geographically remote areas via atmospheric teleconnections. Climate models appear to be sensitive to the land surface because these changes affect the exchange of water, energy, momentum and carbon. Evidence falls into four key groups:

1. It is well established that large-scale changes in key land surface characteristics (albedo, roughness length, water-holding capacity, roots, etc.) lead to changes in climate and changes in the sensitivity of climate to other perturbations.
2. It is now well established that changes in the surface at regional scales (e.g. the Sahel, Amazonia, etc.) influence regional climate.
3. It is less well established that these changes influence global climate in areas remote from the perturbation, but this evidence is gradually becoming more convincing.
4. It is well established that, in order to explain fully the long-term climate changes in global and regional climate (over time scales of tens to hundreds of thousands of years), the feedback between the climate and the biosphere must be included.

Given these findings, it is clearly a necessity to represent the land surface appropriately in climate models.

4. HOW DO WE REPRESENT THE LAND SURFACE IN CLIMATE MODELS ?

4.1. Background

Climate models are designed to simulate climate (i.e. long time scales, years to centuries) at large spatial scales (continents for example). An evaluation of the performance of current climate models (McAvaney *et al*., 2001) indicates that they are reliable at the spatial scales of continents and at seasonal time scales. In order to generate simulations of climate at these scales, climate models represent diurnal time scales for (at present) grid squares of order 200–300 km in size. Although some climate models can use a higher resolution (∼100 km) this does not fundamentally change the challenge of simulating land surface processes. Thus, while there is little evidence that climate model simulations provide reliable information at small time scales (e.g. days), or at spatial scales of individual grid squares, the land surface processes important to climate must be represented at these scales (see Henderson-Sellers and McGuffie, 1987 or Trenberth, 1992 for background on these issues).

An LSM should either explicitly (in terms of representing individual physical processes) or implicitly (in terms of an averaged or aggregated role) represent those processes that influence climate at time scales from about 15 min (approximately the time step of a climate model) through to the time scales at which a given process affects climate. For example, climate model simulations conducted for 10 years may not need to be concerned with ecosystem dynamics, which tend to affect climate on longer time scales (Sellers, 1992). For simulations run for 100–200 years into the future, LSMs may need to include virtually all key processes that affect the exchange of energy, water, momentum and carbon between the surface and the atmosphere.

4.2. First-generation models

The first LSM was implemented by Manabe (1969) into a climate model that intentionally included a simple and idealized distribution of the oceans and continents and did not attempt to represent the seasonal or diurnal cycle. This LSM used a simple energy balance equation, ignoring heat conduction into the soil (a reasonable assumption given the lack of the diurnal or seasonal cycle). Manabe (1969) implemented a globally constant soil depth and water-holding capacity, where evaporation was limited by soil water content below a threshold; if the soil moisture exceeded a prescribed limit, then further precipitation generated runoff. This parameterization of hydrology is commonly called the ‘Manabe bucket model’. Manabe (1969) acknowledged that these are major simplifications, but defended them on the basis that he was exploring climate within a simplified climate modelling framework. Despite the caveats and simplifications, Manabe (1969) was a key step in the representation of land surface processes in climate models. An illustration of the basic conceptual design of this first-generation model is shown in Figure 8.

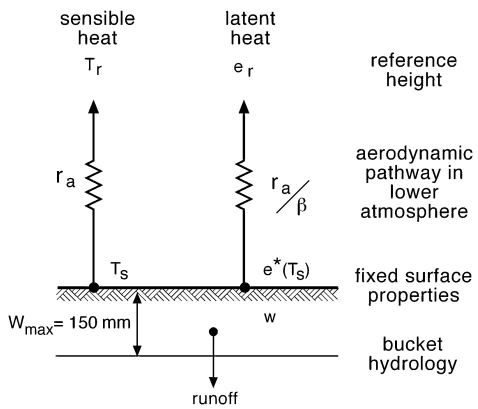

Figure 8. Illustration of a first-generation land surface model. Terms not defined in the text are the reference height for temperature T_r, the maximum soil moisture capacity (W_{max}) and the soil moisture content w. Modified from Sellers *et al*. (1997)

The Manabe (1969) LSM and a generation of simpler schemes were described by Sellers *et al*. (1997) as first-generation models. This is a useful generalization, although Sellers *et al*. (1997) based their classification largely on the level of complexity of the evapotranspiration processes. Their classification recognized that there was a group of LSMs that used simple bulk aerodynamic transfer formulations and tended to use uniform and prescribed surface parameters, including water-holding capacity, albedo and roughness length. Vegetation, and the role of vegetation (e.g. Rind, 1984; Verstraete and Dickinson, 1986) was treated implicitly and did not change in time. It was common to prescribe surface parameters independently (Sellers, 1992), and sometimes poorly (see discussion by Dorman and Sellers, 1989). The uniform water-holding capacity, weaknesses in the representation of evapotranspiration — which did not explicitly include the role of a canopy resistance within the surface resistance (Equation (4)) — and the use of the same aerodynamic resistance for heat, water and momentum were common criticisms (Sellers, 1992; Garratt, 1993). There were many attempts to improve this basic approach by adding limited complexity to these simple schemes, but one of the few systematic attempts to add complexity on the basis that it improved the model performance is very recent (Milly and Shmakin, 2002a,b).

First-generation schemes tended to include a single layer for soil moisture. Manabe (1969) parameterized the available soil moisture by assuming a 15 cm soil-moisture holding capacity globally. Robock *et al*. (1995) have shown that for mid-latitude regions the model works well in comparison with observations and with a more complex model. The Project for the Intercomparison of Landsurface Parameterisation Schemes (PILPS; Henderson-Sellers *et al*., 1995) has shown that the soil moisture simulated by the Manabe (1969) scheme is within the range simulated by other models (e.g. Shao and Henderson-Sellers, 1996). Provided evapotranspiration is calculated properly, the model cannot be distinguished from more complex schemes *at longer time scales* (Desborough, 1999).

The major conceptual limitation of first-generation schemes is the common simplification used to simulate evaporation (see Figure 8):

$$\lambda E = \beta \left(\frac{e^*(T_\mathrm{s}) - e_r}{r_\mathrm{a}} \right) \frac{\rho c_\mathrm{p}}{\gamma} \tag{8}$$

The terms are as in Equation (4), but note that r_s has been omitted and a variable describing the moisture availability β has been added. β ranges between zero (dry) and one (saturated), and this single function attempts to parameterize the full complexity of the various components of the surface resistance term in Equation (4). If the full complexity of the diurnal variability in λE is required from an LSM, or if the function of plants is modelled explicitly, this equation fails to capture the complexity of the real system (see Sato *et al*., 1989). However, while many have argued that this equation is inappropriate because of the need to model more of the processes explicitly (Sellers *et al*., 1997) at the time and space scales appropriate to climate modelling, it is difficult to demonstrate that the use of a simple β-function will necessarily limit climate simulations. It is fair to argue, however, that, overall, the first-generation LSMs did not provide a suitable framework to accommodate the required changes to enable modelling of CO_2 exchange, or to enable experiments to be performed to explore the impacts of LCC.

In addition to simplifications in the evaporation process, first-generation schemes tend to include one to two layers for soil temperature, which may not be adequate to capture the variations in temperature on time scales ranging from the diurnal time scale to multi-annual scales. Runoff was parameterized very simply, with all incident water infiltrating until a critical point, whereafter further rainfall or snow melt becomes runoff. There is good evidence from PILPS that this level of representation of hydrology is inadequate to capture the observed behaviour of hydrological processes (Liang *et al*., 1998; Wood *et al*., 1998; Schlosser *et al*., 2000).

4.3. Second-generation models

4.3.1. Background. A fundamental step forward in land surface modelling occurred in 1978 when Deardorff (1978) introduced a method for simulating soil temperature and moisture in two layers and vegetation as a single bulk layer. Deardorff's soil model built heavily on the 'force–restore' soil schemes by Bhumralkar (1975) and Blackadar (1976) (also, see Deardorff, 1977). In terms of vegetation, Deardorff (1978) proposed a single layer of vegetation that shielded a fraction of the ground from solar radiation. The representation of heat and moisture transfer from the canopy built heavily on the micrometeorological work of Legg and Long (1975) and Thom (1972) and, overall, the model implemented the philosophy of Monteith and Szeicz (1962) for evaporation. In effect, Deardorff (1978) brought together a collection of work and packaged it into something that was recognizable as a physically based LSM.

Sellers *et al*. (1997) described those LSMs that included a Deardorff (1978)-type model of vegetation being as 'second generation' (see Figure 9). Deardorff (1978) represented a revolution in land surface modelling, since processes were treated explicitly and mathematically and this provided an opportunity for a generation of micrometeorologists to contribute to LSM development. The two key players (who, it turned out, did not come from a micrometeorology background) were R. E. Dickinson and P. J. Sellers, who developed LSMs that built philosophically on Deardorff (1978). Key papers included those by Dickinson (1983, 1984), Dickinson *et al*. (1986, 1993), and Sellers *et al*. (1986, 1992b, 1994, 1996). The two LSMs, the Biosphere Atmosphere Transfer Scheme (BATS) and the Simple Biosphere Model (SiB), continue to be developed and form the basis for some major new innovations, such as the Common Land Model (Dai *et al*., 2002).

There are a very large number of second-generation models that are innovative in the way some components have been developed or tested, but all are fundamentally built from the leadership of Deardorff, Dickinson and Sellers. Examples include Abramopoulos *et al*. (1988), Noilhan and Planton (1989), Koster and Eagleson (1990), Xue *et al*. (1991), Wood *et al*. (1992), Verseghy *et al*. (1993), Ducoudre *et al*. (1993), Thompson and Pollard (1995), Viterbo and Beljaars (1995), Wetzel and Boone (1995) and Desborough and Pitman (1998). Many of these LSMs continue to be developed via improvements in scheme components, data input, computational efficiency, etc. This generation of LSMs underpins virtually all work performed to date on the sensitivity of climate models to surface characteristics, including deforestation experiments, desertification experiments and work focusing on the impact of LCC in general (discussed earlier). These LSMs also provide the basis for almost all recent work conducted on the impact of increasing CO_2 on climate (McAvaney *et al*., 2001).

4.3.2. Evaporation in second-generation schemes. These second-generation models vary in detail, but they have many components in common. As discussed by Sellers *et al*. (1997), second-generation models usually

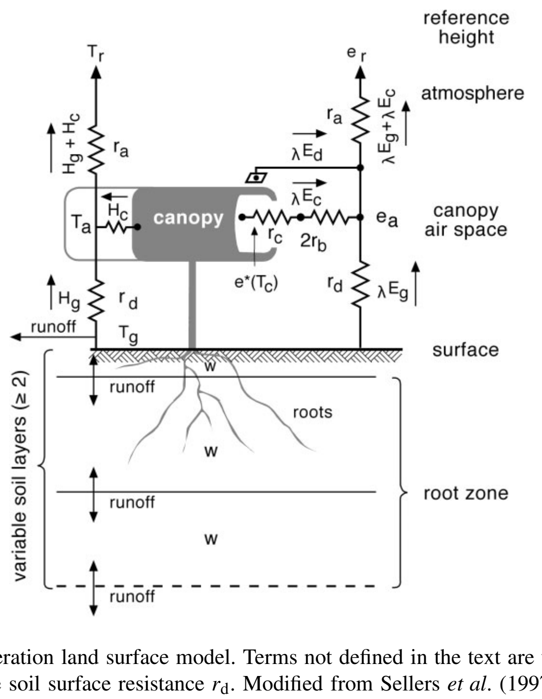

Figure 9. Illustration of a second-generation land surface model. Terms not defined in the text are the boundary-layer resistance r_b, and the soil surface resistance r_d. Modified from Sellers *et al*. (1997)

represent the vegetation–soil system such that the surface interacts with the atmosphere, rather than being passive (as in the first-generation models). The second-generation models differentiate between soil and vegetation at the surface; thus, albedo may vary spatially across a grid square, as well as varying depending on the wavelength of the incoming solar radiation. Canopies are highly effective at absorbing in wavelengths of 0.4–0.7 μm (the photosynthetically active radiation (PAR)) and are moderately reflective in the near-infrared (0.72–4 μm) (e.g. Dickinson, 1983; Dickinson *et al.*, 1987). This difference, captured in second-generation models, provided a major opportunity to begin integrating satellite data into LSMs (e.g. Sellers *et al.*, 1994). Second-generation LSMs also explicitly represented the impact of vegetation on momentum transfer. Canopies are rough and generate turbulence, which enhances the exchange of H and λE, and capturing this enhancement was an important step forward. These first two characteristics of second-generation models could, in principle, have been added to first-generation schemes. The third characteristic could not have been, since it required an explicit Deardorff (1978)-style canopy architecture. Second-generation schemes include, in some form, an explicit biophysical control on evaporation. Plants have evolved to regulate water use to maximize their ability to fix carbon via photosynthesis. The process of photosynthesis, where PAR fuels the combining of CO_2 and water into sugars and other organic compounds, required CO_2 to be transferred from the atmosphere through stomates on the leaf surface. When stomates are open to allow CO_2 uptake they are also open to water vapour molecules, which are lost to the atmosphere.

The community of plant physiologists, a very different community to those developing LSMs for climate purposes, have researched the mechanism of CO_2 and water vapour exchange between leaves, through stomates, into the atmosphere for many years. Some empirical work suggested that the conductance of water through stomates, controlled by the stomatal conductance g_{st} (the inverse of which is the stomatal resistance r_{st}), was described by a simple equation:

$$g_{st} = \frac{1}{r_{st}} = g_{st}\ (\mathrm{PAR})[f(\delta e)f(T)f(\psi_l)] \tag{9}$$

This relationship, developed by Jarvis (1976), was used in most second-generation schemes and captures the key responses of stomates to PAR, humidity δe and temperature T. It also includes the leaf water potential ψ_l, which can be related to soil water content in the root zone, root distribution and evaporative demand. In effect, a formulation similar to Equation (9) was included by Deardorff (1978) based on a range of meteorological observations (e.g. Monteith and Szeicz, 1962). A major effort through the 1980s and 1990s investigated the form of each of the functions in Equation (9) (Schulze *et al.*, 1994), and models that use this formulation could behave differently simply as a result of the precise implementation of each functional form.

If the stomatal conductance is modelled explicitly, then (modified from Sellers *et al.*, 1997):

$$\lambda E = \frac{e^*(T_s) - e_r}{r_c + r_a}\frac{\rho c_p}{\gamma} \tag{10}$$

where most terms are as described in Equation (4), but note that r_c in Equation (10) replaces the surface resistance r_s of Equation (4). Thus, Equation (10) only simulates the evaporative flux from within the leaves to the atmosphere, whereas Equation (4) represents the total evaporation from the surface. In implementing Equation (10) into an LSM, the fluxes of water from the soil (H_g, λE_g) and from intercepted precipitation on the canopy (λE_d) must be represented separately (Figure 9). Also note that the canopy resistance r_c is the inverse of the canopy conductance g_c and that these two terms can be obtained from the stomatal conductance g_{st} or stomatal resistance r_{st} via appropriate scaling from a stomatal-scale function to a canopy-scale function. A common method (Dickinson *et al.*, 1998) is to use LAI thus:

$$r_c = r_{st}/\mathrm{LAI} \quad \text{or} \quad g_c = g_{st}\ \mathrm{LAI} \tag{11}$$

Equation (10) differs fundamentally from its equivalent in a first-generation LSM, which omits any explicit consideration of r_c, and represents the total resistance to evaporation from the soil and canopy system using a single β function (e.g. Equation (8) and Figure 8). The inclusion in second-generation schemes of r_c in parallel with r_a (Figure 9) realistically separates aerodynamic and canopy-related resistance terms. Since these models also differentiate between soil and vegetation (Figure 9), a separate resistance to evaporation from the soil has to be incorporated, but this can be analogous to the β-function in Equation (8). In almost all cases, the combination of these changes reduces evaporation from that simulated by first-generation LSMs.

4.3.3. Hydrology in second-generation schemes. Second-generation schemes model interception by the canopy, although it is possible to include this process in a first-generation scheme (Desborough, 1999). Second-generation schemes also tend to add more complexity into the soil moisture parameterizations, replacing the simple soil hydrology of Manabe (1969) by methods based on the implementation of the Richards (1931) equation and the development of soil physics by Philip (1957) and Hillel (1982). The vertical transfer of water within the soil profile (*Q)* in second-generation schemes usually uses a set of diffusion equations based on Darcy's law:

$$Q = K\left(\frac{\partial\psi}{\partial z} + 1\right) \tag{12}$$

where ψ is the soil moisture potential (soil suction), K is the soil hydraulic conductivity, and z is depth. To close this equation, it is necessary to assume forms for the hydraulic conductivity and the soil moisture potential as a function of soil moisture. The dependencies provided by Clapp and Hornberger (1978) are commonly used in climate models, largely due to their simplicity:

$$K = K_s W^{2b+3} \tag{13}$$

$$\psi = \psi_s W^{-b} \tag{14}$$

where K_s is the saturated hydraulic conductivity, W is the soil wetness fraction, ψ_s is the value of the soil moisture potential at saturation and b is an empirical constant. Cosby *et al.* (1984) provide a means to link these dependencies to soil properties. More complex forms of these functions exist and are used in some LSMs (e.g. Van Genuchten *et al.*, 1991), and Viterbo and Beljaars (1995) discuss the implementation of these relationships in more detail. Global data sets for parameters K_s, ψ_s, and b must be available for several soil texture classes in order to use these relationships. It is usual to ignore variation in these parameters with soil depth, given the scarcity of data.

Surface runoff can be parameterized very simply (all incident water infiltrates until saturation and all remaining water becomes runoff) or in complex ways (based on a parameterization of the rate at which water infiltrates into the soil), or in a wide variety of ways in between, including calculating infiltration as the residual of the surface water balance (see Wetzel *et al.*, 1996). Unless routing is employed, the surface runoff is assumed to be lost into the ocean *immediately*. This appears to be a gross simplification, but at the spatial scales of climate models, and for the purposes of simulating climate, this was reasonable until runoff began to be used as a boundary condition for ocean models. Some LSMs now include river routing (e.g. Sausen *et al.*, 1994; Hagemann and Dümenil, 1998) in order to simulate the annual cycle of river discharge into the ocean, and this appears to improve the modelling of runoff from some large drainage basins (Dümenil *et al.*, 1997), although water storage and runoff in regions of frozen soil moisture remain outstanding problems (Arpe *et al.*, 1997; Pitman *et al.*, 1999). The inclusion of routing also assists in the use of runoff for validating LSMs at large spatial scales. Vörösmarty *et al.* (1989) and Liston *et al.* (1994) have independently developed river routing schemes for use within climate models based on a linear reservoir, and Miller *et al.* (1994) have developed a model to be applied globally. A recent advance by Warrach *et al.* (2002) demonstrates the advantages of including topographic controls on runoff. These are all useful developments, but weaknesses remain and these are mainly linked to difficulties in parameter estimation (Todini and Dümenil, 1999).

Major uncertainty remains in our ability to model the surface water balance. Chen *et al.* (1997), Wetzel *et al.* (1996), Liang *et al.* (1998), Pitman *et al.* (1999), and many others have demonstrated the strong interaction between the surface water balance and the surface energy balance, indicating that systematic errors in the modelling of runoff lead inevitably to systematic errors in the modelling of the partitioning of available energy between λE and H, and this impacts on the simulation of climate. Some of this uncertainty is driven by the parameterization of processes that generate runoff and the differences in the storage characteristics of the models (Gedney *et al*., 2000), but, overall, difficulties in modelling hydrological processes provide a very strong rationale for improving the simulation of hydrology in climate models.

A recent major philosophical advance for an LSM designed for climate models is the evolution of a more catchment-based model (Koster *et al*., 2000). This LSM allows for a catchment-based framework where different parts of the catchment can represent different hydrological regimes. The viability of the model has been demonstrated (Ducharne *et al*., 2000) and the results are similar to those generated by more traditional LSMs. However, the philosophical change from an atmospherically defined grid to a catchment-based framework would seem to offer a route for long-term progress over the next decade, provided that the links between the catchment-based LSM and the grid-based atmospheric model can be made efficient and robust.

4.3.4. Snow in second generation models. The first significant effort in linking a reasonably sophisticated snow sub-model into a climate model was probably by Hansen *et al*. (1983). Models such as BATS (Dickinson *et al*., 1986, 1993) and SiB (Sellers *et al*., 1986) parameterized snow as part of the upper soil layer for thermal processes and as a separate layer for hydrological processes. A later scheme, CLASS (Verseghy *et al*., 1993) simulated snow as a discrete layer for both thermal and hydrological processes. Recent developments in LSMs have begun to pay more attention to snow. For instance, Loth and Graf (1993) developed a dedicated two-layer snow scheme, and more complex snow parameterizations were developed by Lynch-Steiglitz (1994), Thompson and Pollard (1995) and Douville *et al*. (1995a,b).

Work on testing and evaluating snow schemes has also been quite extensive. Individual efforts (e.g. Douville *et al*., 1995a; Yang *et al*., 1997) have been supplemented by major intercomparisons of snow in LSMs (Schlosser *et al*., 2000; Slater *et al*., 2001; and the Snow Model Intercomparison Project, SNOWMIP, http://www.cnrm.meteo.fr/snowmip/). There have also been analyses of the performance of LSMs coupled within the climate models. This is extremely difficult due to the feedbacks and variability within the coupled system, but useful results have been provided by, for example, Foster *et al*. (1996) and Frei and Robinson (1998).

A major new assessment of the ability of LSMs to model snow has been performed by Bowling *et al*. (in press). PILPS intercomparison results from Scandinavia showed that LSMs performed generally quite well, capturing the effects of low net radiation. Importantly, the results showed that, although summer evaporation was highest, winter sublimation rates had a proportionately greater influence on annual runoff, which Nijssen *et al*. (in press) attributed to the way snow surface roughness was parameterized. Overall, the way in which snow is modelled and the way in which snow–vegetation–hydrology feedbacks are captured (e.g. Betts, 2000) deserve more attention and remain significant limitations in current LSMs.

4.3.5. Summary. Second-generation models improved the physical representation of the continental surface and probably the simulation of climate because, along with improvements in the simulation of evaporation, vegetation parameters, etc., they also contained improved soil temperature and soil moisture representations. They certainly provided a means to explore the impact of LCC (see Section 3.2). It is an open question as to what extent changing a first-generation scheme to a second-generation scheme really improved the simulation of climate. The change between schemes necessitates a change in model physics, model parameters and often changes in the host climate model. Although it is not always absolutely clear where subsequent improvements in climate simulations originate from, Desborough *et al*. (2001) showed that, as the complexity of the surface energy balance increases in an LSM, the climate does appear to be affected. The recent examination of the impact of adding complexity into an LSM on model performance (Milly and Shmakin, 2002a,b) is a rare example of step-by-step LSM development driven by a systematic assessment of the relationship between additional complexity and model performance.

Overall, there is a definite majority view that second-generation schemes perform better than first-generation schemes. In a suite of model intercomparison studies conducted via PILPS, the range of second-generation schemes usually outperforms the simpler schemes (Chen *et al*., 1997; Pitman *et al*., 1999; Schlosser *et al*., 2000; Bowling *et al*., in press). However, demonstrating improvements in performance *within* the range of second-generation schemes resulting from the modelling of a particular process, for example, in a particular way, has proved problematic.

Finally, there is evidence strongly suggesting that the second-generation models do improve the modelling of surface–atmospheric exchanges, at least on the time scale of days. Work by Beljaars *et al*. (1996) showed an improvement in a precipitation weather forecast as a result of improved surface processes. Viterbo *et al*. (1999) have provided further evidence supporting the need for good models of soil processes to simulate European surface temperatures (in winter). This evidence, although sourced from the weather forecasting community at spatial and temporal scales more appropriate to that science, is very supportive of the community's move to second-generation schemes.

4.4. Third-generation LSM

The major advance, and major limitation, of second-generation LSMs is that they model canopy conductance empirically, taking into account plant and environmental conditions, but they only use this conductance to model transpiration. It was recognized in the late 1980s that the addition of an explicit canopy conductance provided a means to improve the simulation of the evapotranspiration pathway, as well as to address the issue of carbon uptake by plants. The 'greening' of LSMs represented a major revolution in our modelling capability.

The addition of carbon into LSMs needed the support of the plant physiology community. They had established that the leaf assimilation of carbon was limited by the efficiency of the photosynthetic enzyme system (Rubisco-limitation), the amount of PAR captured by the leaf chlorophyll and the capacity of the leaf to utilize the products of photosynthesis. The integration of this photosynthesis capability into LSMs probably began through a cross-fertilization of communities during some major fieldwork campaigns. The most well-known campaigns are the HAPEX–MOBILHY (André *et al*., 1988), FIFE (Sellers *et al*., 1992a), HAPEX–Sahel (Goutourbe *et al*., 1997), and ABRACOS (Gash *et al*., 1996) experiments. The FIFE experiment brought together communities that included plant physiologists, micrometeorologists and hydrologists under the International Satellite Land Surface Climatology Project (ISLSCP), a group then led by P.J. Sellers.

The cross-fertilization of ideas seems to have been effective and, soon after FIFE, Collatz *et al*. (1991) and Sellers *et al*. (1992b) began to integrate models of stomatal conductance and photosynthesis into LSMs building on work by, for example, Farquhar *et al*. (1980) and Farquhar and von Caemmerer (1982); also, see Leuning (1995). This effort changed the level of representation of canopy conductance in LSMs. Instead of representing canopy conductance empirically (Equation (9); Jarvis, 1976), it became represented via semi-mechanistic models of leaf photosynthesis based on how stomates are believed to function (Figure 10). The combination of the evidence of limitations on leaf assimilation of carbon and the knowledge that stomates function to maximize the efficiency of plant water use (e.g. Cowan and Farquhar, 1977; Ball *et al*., 1987; Ball, 1988) permitted a semi-empirical model of leaf conductance to be proposed (Collatz *et al*., 1991):

$$g_s = m\frac{A_n h_s}{c_s} + b \tag{15}$$

where A_n is the net leaf assimilation rate, c_s is the partial pressure of CO_2 at the leaf surface, h_s is the relative humidity at the leaf surface, m is an empirical coefficient ($m \approx 9$ for most C_3 plants and $m \approx 4$ for C_4 plants) and b is the minimum stomatal conductance (around 0.01 for C_3 and 0.04 for C_4 plants); see Sellers (1992) and Sellers *et al*. (1997). Predicting the stomatal conductance requires A_n to be derived. Simplifying from Collatz *et al*. (1991):

$$A_n = \min\left\{\begin{array}{l}\text{Light limitation}\\ \text{Rubisco limitation}\\ \text{Capacity utilization limitation}\end{array}\right\} - R_d \tag{16}$$

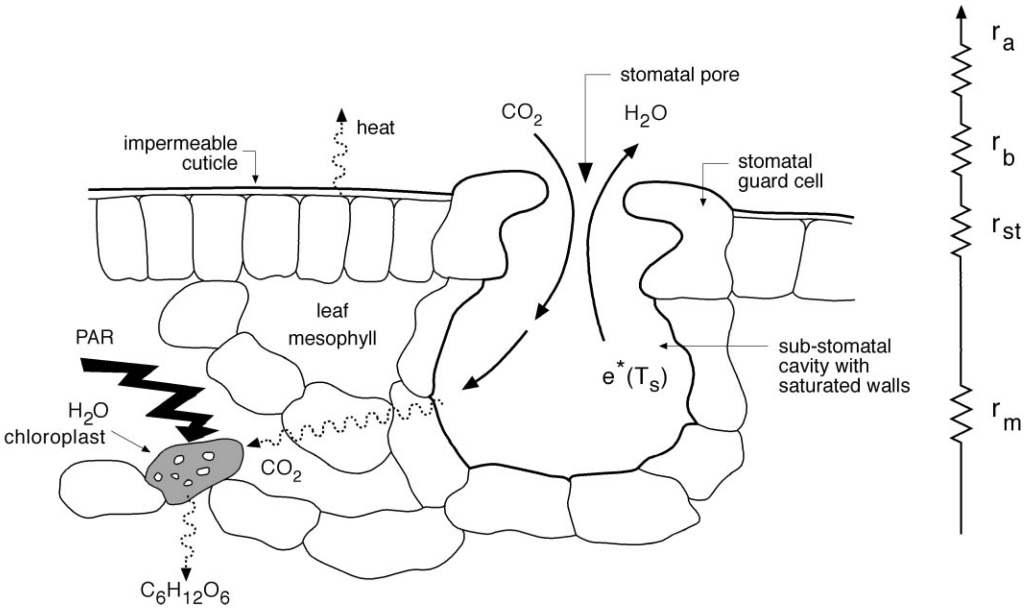

Figure 10. Schematic of a cross-section of a leaf. The resistance not defined in the text is the mesophyll resistance (r_m)

where R_d is the leaf respiration rate. Collatz *et al.* (1991) explain how to solve this equation in some detail, and there is also some useful discussion in Bonan (1995) and Cox *et al.* (1998). Overall, Equations (15) and (16) represent a coupled model of stomatal conductance and photosynthesis. By assuming that mean incident PAR and leaf nitrogen concentrations were proportional through the plant canopy, Sellers *et al.* (1992b) derived methods to scale these leaf-level models to canopy-scale models, giving estimates of the canopy resistance r_c, canopy photosynthesis A_C and canopy respiration R_D shown in Figure 11. From these, the gross primary productivity can be obtained Π_g (Cox *et al.*, 1998):

$$\Pi_g = 0.012(A_C + R_D\beta) \tag{17}$$

where β is a moisture limitation term imposed on the dark respiration rate and the 0.012 factor converts from units of mol CO_2 m^{-2} s^{-1} to kg C m^{-2} s^{-1}. Net primary productivity Π is

$$\Pi = \Pi_g - R_p \tag{18}$$

where R_p is the plant respiration, split into maintenance and growth respiration. This can be allocated in a variety of ways to influence vegetation phenology (Figure 12), but it needs knowledge of the nitrogen distribution. This coupling of the transpiration and photosynthesis parameterizations brought biology and biochemistry into the heart of LSMs (Figure 11). It also brought about the potential to model the role of the biosphere explicitly in LSMs and, most critically, the biospheric sink for CO_2. Key individuals, including Sellers *et al.* (1992b), Bonan (1995) and Cox *et al.* (1998), built this capacity into LSMs and used them within climate models and provide reasonably accessible descriptions of the overall procedure for modelling the link between photosynthesis and canopy conductance (Figure 11).

Once net carbon assimilation or net primary productivity has been calculated, an opportunity then exists to do something with the assimilated carbon. In nature, plants use a net accumulation of carbon to grow leaves, branches, roots, etc. This is a net loss of CO_2 from the atmosphere, which constitutes the terrestrial carbon sink. Taking a net accumulation of carbon and partitioning it such that growth occurs is not a traditional

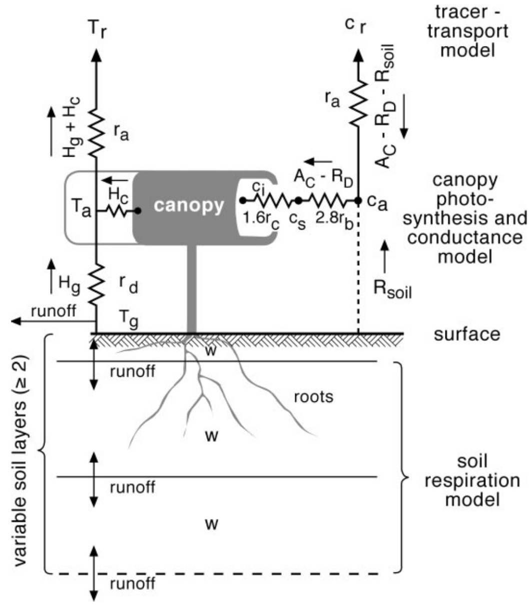

Figure 11. Illustration of a third-generation land surface model. Terms not defined in the text are partial pressures of the canopy air space c_a, internal to the leaf c_i, at the leaf surface c_s and at a reference height c_r. A_c is the canopy photosynthesis rate, R_D is the canopy respiration rate and R_{soil} is the soil respiration rate

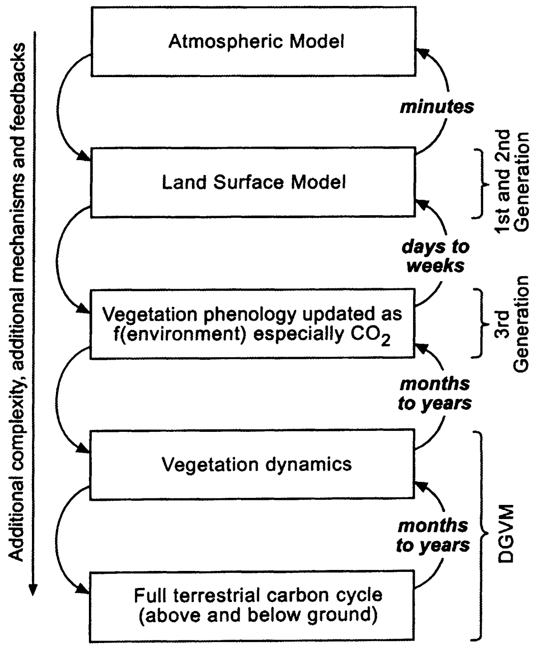

Figure 12. Schematic of the increasing levels of detail being added into surface modelling approaches. The second box represents first- and second-generation land surface models. The addition of vegetation phenology (using via semi-mechanistic models of leaf photosynthesis and respiration) defines third-generation models. The allocation of the net carbon balance, and other additions to reflect the full terrestrial carbon cycle translates a 'land surface scheme' into a dynamic global vegetation model DGVM. In each case, the requirements are additive, so that a fully coupled DGVM requires the traditional (usually second-generation) land surface model

area of expertise in land surface modelling (which is limited to the top two boxes of Figure 12), but it is within the ecological community. The simplest addition to a typical LSM resulting from the knowledge of net CO_2 assimilation is to allow the leaves to respond, and thus include the structural feedback of vegetation (third box, Figure 12). Dickinson *et al*. (1998) allocate carbon to leaves, convert it to carbon assimilation per unit leaf area and thus allow leaves to grow. They also allow for root and wood allocation and use a simple soil carbon model based on Parton *et al*. (1987). Thus, an LSM that has been able to respond to changes in climate through influencing energy and water exchange (although with largely static vegetation-related characteristics) can now respond in two further ways to a climate change. It can now respond physiologically as increasing CO_2 influences the canopy conductance and it can respond structurally by growing different leaves or taller trees (Figure 7). Evidence already exists to show that representing these two feedbacks is important (Henderson-Sellers *et al*., 1995; Pollard and Thompson, 1995; Betts *et al*., 1997; Levis *et al*., 2000; Bergengren *et al*., 2001). Overall, the addition of these processes represents a fundamental advance in LSMs towards a realistic representation of significant feedbacks missing in climate simulations of increasing CO_2: the response of the biosphere.

Thus, third-generation schemes are identifiable by the method used to model carbon. These LSMs tend to employ representations of other processes (soil temperature, soil hydrology, runoff, etc.) that are similar to those included in second-generation LSMs (Figure 11).

As land-surface modellers have built the LSMs, they have observed the parallel development of a suite of ecological models by a different community. These ecological models tended to focus on carbon and other biogeochemical cycles (bottom two boxes of Figure 12), use plant functional relationships to categorize the vegetation (Bonan *et al*., 2002) and use time steps that made linking them into climate models problematic (Martin, 1993). These ecological models tended to focus on how the terrestrial biosphere responded to the atmosphere (on time scales of months to years) rather than how the land surface partitioned energy and water as a boundary condition for the atmosphere.

As the role of carbon in climate became central to climate modelling (and policy development), and as land-surface modellers moved into second- and towards third-generation schemes, and as the ecological community recognized that their understanding and skills were valuable to the climate modelling community, the potential for a merging of traditional LSMs with the ecological models increased. Since climate models *require* land surface inputs every 15–20 min, aspects of a traditional LSM must be retained, but it is possible now to link such a model to an ecological model asynchronously (Figure 12) and begin to address the full detail of the terrestrial carbon cycle. The Cox *et al*. (1999) model is an example of a group known as DGVMs. Other examples have been developed by Foley *et al*. (1996), Haxeltine *et al*. (1996) and Tian *et al*. (1999). Intercomparisons and discussions of these terrestrial ecosystem models are provided by McGuire *et al*. (2001) and Cramer *et al*. (2001).

The merging of the ecological approaches with LSMs (i.e. the full complexity of Figure 12) is extremely challenging, but it opens up major opportunities to link the physical and biophysical sciences and provides tools to begin to address key questions regarding the future of global-scale biogeochemical cycles. These developments will be discussed further in Section 5.

5. KEY FUTURE DEVELOPMENTS IN CLIMATE MODELS

Although the role of the land surface in climate, through its influence on the exchange of energy, water, carbon and momentum, has been known for many years, it took almost 20 years for the representation of these processes in climate models to evolve from the simple approach of Manabe (1969) to the more physically based approach of Dickinson *et al*. (1986) and Sellers *et al*. (1986). These more biophysically based models now form the basis for virtually all our understanding of:

- the role of the land surface in simulating climate and climate variability in climate models;
- the impact of changes in the land surface processes on climate; and
- the role of the land surface in climate change and climate sensitivity.

These biophysically based models (the second-generation schemes in Sellers *et al.*'s (1997) terminology) are remarkable in their capacity to capture the complexity of energy and moisture exchange within the soil, vegetation and snow system, and they can also be linked to hydrological models to capture catchment-scale controls on hydrological processes (e.g. Koster *et al.*, 2000). There is little evidence that the simulation of the current climate, or the simulation of current climate variability, is substantially limited by weaknesses in these second-generation models. Thus, if the focus of scientific enquiry is related to decadal-scale climate simulations or shorter, these second-generation schemes are most likely a suitable means by which to represent land surface processes. However, the need for an LSM to capture other processes is increasing, in particular the role of the surface within various biogeochemical cycles and within the hydrological cycle. These areas provide foci for future developments.

5.1. Carbon

The main terrestrial sink for increasing atmospheric CO_2 is via plants, and if plants take up more carbon then they will respond physiologically (changes in canopy conductance), structurally (larger leaves, longer roots, etc., Figure 7), and biogeographically (involving changes in vegetation type and distribution). Climate model simulations of LCC or the impact of increasing CO_2 on climate appear to be sensitive to how this biospheric feedback is included in LSMs. Betts (2000), for example, has shown that both carbon storage and biophysical feedbacks need to be accounted for when considering reafforestation for climate-change mitigation (see Pielke *et al.*, 2002 for further discussion). Bounoua *et al.* (1999) showed that the physiological effect of reduced stomatal conductance resulting from increased CO_2 and the radiative effects of increased CO_2 interacted in complex ways. Although this area remains one of conjecture, it is clear that a failure to represent these processes in climate models has the capacity to bias climate-model projections of climate over the next 100–200 years.

Perhaps the most dramatic recent finding of the role of the biosphere in feedbacks over the next century was provided in papers by Cox *et al.* (2000) and Friedlingstein *et al.* (2001). These two groups used independently developed climate models coupled to third-generation LSMs. The Friedlingstein *et al.* (2001) group used a terrestrial carbon model developed by Friedlingstein *et al.* (1995) and Ciais *et al.* (1999) that is based on well-established ecology (Field *et al.*, 1995). The Cox *et al.* (2000) experiments used an LSM developed by Cox *et al.* (1998, 1999) based, in part, on Collatz *et al.* (1991, 1992). The model allows a dynamic evolution of land cover based on a Lotka–Volterra-type approach, assumes that the soil carbon respiration rate doubles with each 10 °C of warming (Raich and Schlesinger, 1992) and includes a dependence on soil moisture. There is some support for the soil–carbon–temperature relationship (Jones and Cox, 2001) and some evidence that it is too simplistic (Glardina and Ryan, 2000).

Both groups found that, as the Earth warms due to increasing CO_2, the capacity of the terrestrial biosphere to absorb and store carbon declines. Cox *et al.* (2000) find that the terrestrial biosphere functions as a sink to about 2050 and then turns into a source (Figure 13). While vegetation continues to take up CO_2 past 2050, the rate is reduced and is overwhelmed by the collapse of the soil-carbon sink, which then begins to release vast amounts of carbon into the atmosphere (Figure 13). Friedlingstein *et al.* (2001) find that the size of the sink increases at first, but then declines as temperature increases. In a major intercomparison exercise, Cramer *et al.* (2001) compared results from six of these global biospheric models to prescribed increases in CO_2 (these were not experiments performed coupled with a climate model). The results, as well as the coupled results from Cox *et al.* (2000) and Friedlingstein *et al.* (2001), are different in detail, but they all demonstrate an important but uncertain role for the future biosphere. Assessments of future CO_2-related warming have not factored in the considerable amplification that would occur if the Cox *et al.* (2000) results are correct.

Increasing CO_2 also affects radiative forcing and the energy available for land surface processes. Second-generation schemes adapt to this increase in energy through physiological changes in the stomatal conductance, and the first-order biospheric feedbacks of this increased radiative forcing on the exchange of latent and sensible heat can therefore be modelled. Second-generation schemes probably capture the impact of increases in radiative forcing following changes in CO_2 quite reasonably. However, the increase in CO_2 also directly

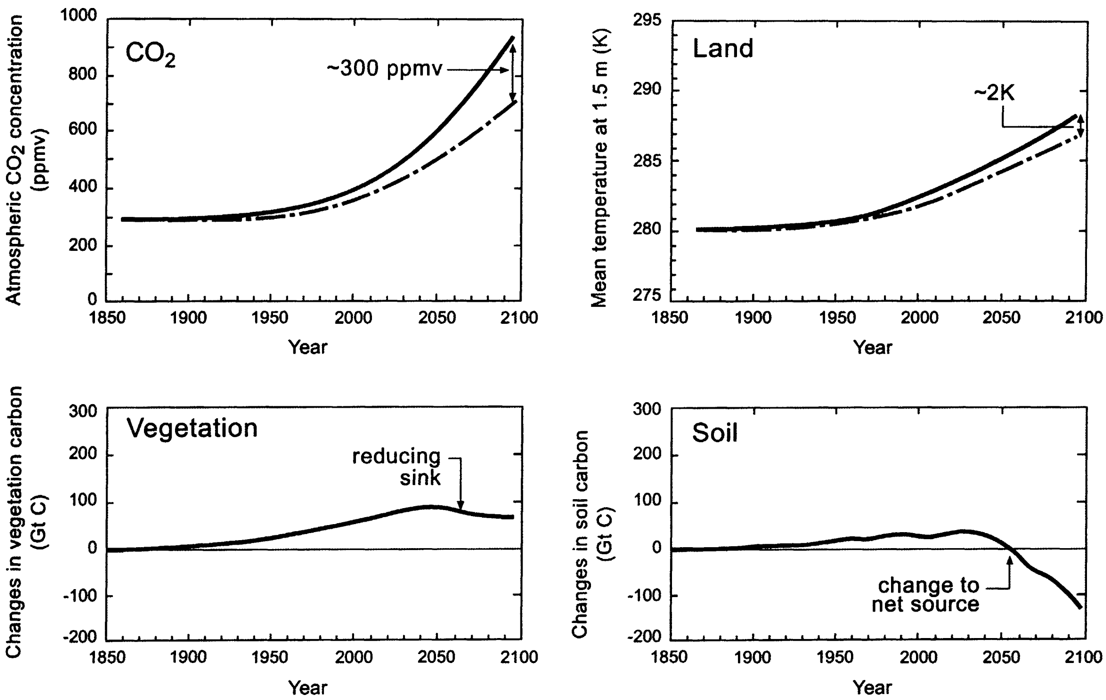

Figure 13. The impact of the biospheric feedback on future climate change. The first panel (CO_2) shows the difference in CO_2 concentration between 1850 and 2100 resulting from the impact of warming on the carbon cycle (solid line shows the change in CO_2 including the biospheric feedback; the dashed line is the change without this feedback). The difference in CO_2 is associated with more warming if the biospheric feedback is included (second panel). The third panel shows the change in the carbon sink into the vegetation (reducing since from ~2050) and the final panel shows the collapse of the soil carbon sink from ~2050 as a result of warmer temperatures. After Cox *et al.* (2000)

affects biospheric activity (Mooney *et al.*, 1991; Stitt, 1991; Walther *et al.*, 2002), and second-generation schemes cannot capture the impact of increasing CO_2 on stomatal conductance cannot model the uptake of CO_2 by the biosphere, and cannot respond structurally to increases in this uptake. This is a crucial limitation, since the biospheric sink for CO_2 is very substantial (Prentice *et al.*, 2001): it may remain in a steady state over the next century, it may increase, or it may decline. Projections for the IPCC (Houghton *et al.*, 1996, 2001) assume, implicitly, that the terrestrial carbon sink continues to take up a fixed fraction of anthropogenic emissions. However, small perturbations in this sink may have major implications for the rate of increase of CO_2 in the atmosphere (Pielke *et al.*, 1998). If the Cox *et al.* (2000) results prove correct, then the weakening of the net biospheric sink as temperatures increase leads, by 2100, to CO_2 levels nearly 300 ppmv higher and ~2.0 °C extra warming between 1860 and 2100 (Figure 13), compared with a standard IPCC scenario (IS92a). The issue of how the land surface responds to increasing CO_2 is therefore very relevant to key issues like global warming. However, the biospheric feedback to LCC experiments is also potentially significant, and efforts are just beginning in these LCC experiments to explore the combined role of surface change in an environment of increasing CO_2 (Costa and Foley, 2000; Bergengren *et al.*, 2001; Zhang *et al.*, 2001).

Simulating the interactions between changes in CO_2 and biospheric activity requires a third-generation scheme. There are very few third-generation schemes coupled into climate models and available to explore questions related to climate and climate change. However, efforts to increase the number of third-generation schemes coupled to climate models are under way, and it is probable that they will replace the second-generation schemes over the next 5 years. This provides the opportunity to repeat a vast range of experiments conducted with second-generation schemes to determine whether biospheric feedbacks affect the conclusions reached to date on the role of land surface processes in climate models.

5.2. Other major developments

While the addition of full carbon capability provides an exciting enhancement of the capability of LSMs, the modelling of these processes requires further improvements in other components of these schemes. Typical second- and third-generation schemes contain multilevel soil temperature and moisture schemes, perhaps with routing of runoff, combined with a multilevel snow parameterization. Every single component of these models requires development. In effect, LSMs need to develop such that the individual components are balanced, not necessarily in terms of their complexity but in terms of the role they play in affecting surface–atmospheric exchanges. A common criticism of SiB (Sellers *et al.*, 1986), for example, was that its above-ground processes were modelled in a complex way, but below-ground processes were far more simplistic. These two components are linked via root-uptake, and errors in the hydrological parameterization inevitably cause errors in the surface energy balance simulation, which outweigh gains made via complex above-ground parameterizations. Thus, the movement towards third-generation schemes and carbon has to be balanced with improvements in the capability of the below-ground components of models.

Although many areas could be highlighted here, there are four that seem of particular concern. First, in the hydrological components (soil moisture and runoff), LSMs typically use one-dimensional mass conservation equations based on Darcy's law. Thus, processes such as drainage, surface runoff and movement of water within the soil are assumed to occur only vertically. Account is not always taken of the basic principles of hydrology (see Beven (2000)) or the principle of the catchment as the basic unit within which hydrological processes operate. Some efforts in this direction (to make second-generation schemes more hydrologically focused) have been made (Wood *et al.*, 1992; Liang *et al.*, 1994), but the Koster *et al.* (2000) approach discussed earlier appears to provide a genuine opportunity to advance this area of land surface processes. It is important that this is pursued, given that there is increasing evidence that the failure to capture the hydrological processes well limits our capacity to model future climate change (Crossley *et al.*, 2000; Gedney *et al.*, 2000). Indeed, it appears that the ways in which the hydrology are modelled may be more important than the surface energy balance, at least in second-generation schemes (Pitman and McAvaney, 2002). Other issues relating to hydrology, including permafrost, the impact of frozen soils on infiltration and the whole issue of groundwater, are areas that LSMs have hardly begun to consider, despite their potential importance.

A second area of concern is that higher CO_2 concentrations may lead to changes in root distribution (Norby and Jackson, 2000), which is expected to affect the pattern (in time and space) of evaporation (de Rosnay and Polcher, 1998; Kleidon and Heimann, 1998). Increasing carbon assimilation may increase root growth, but increased temperature also increases root mortality, and generalizations are difficult, since plants have evolved to cope with a wide range of air and soil temperatures. The whole issue of roots in global change is simply problematic (Norby and Jackson, 2000), despite relatively good observational databases (Canadell *et al.*, 1996; Jackson *et al.*, 1996). We know that LSMs are sensitive to roots (Desborough, 1997; Kleidon and Heimann, 1998), but it is not well established how best to represent roots, root water uptake and the allocation of carbon for root growth in LSMs (Feddes *et al.*, 2001). Some changes in roots affect moisture availability and may act as a feedback on an initial perturbation, and this area deserves substantial attention.

The third major concern is the issue of heterogeneity and sub-grid-scale processes. The land surface is *highly* variable in all aspects in space, such that measurements taken a metre or two apart may report substantial differences in everything from soil moisture, through soil characteristics, to the type of vegetation. Since climate model grid squares are, by necessity, large (around 300×300 km^2 at present), the nature of the interaction between the surface and the atmosphere varies greatly over that area. Various ways have been proposed to handle this variability. There are various ways to aggregate parameters (Chehbouni *et al.*, 1995; Shuttleworth *et al.*, 1997), and there is evidence that the way parameter values are aggregated may matter to climate and climate models (Ek and Cuenca, 1994; Burke *et al.*, 2000; Sen *et al.*, 2001). Recent efforts to explore methods of calibration to provide parameters for LSMs (e.g. Gupta *et al.*, 1999; Xia *et al.*, 2002) also look promising.

The issue of heterogeneity in surface processes in general can be addressed in a variety of ways. One method is the 'statistical–dynamical approach' (see Bonan *et al.* (1993)), where important features of the land surface are represented by a probability density function (Avissar and Pielke, 1989; Entekhabi and Eagleson, 1989; Famiglietti and Wood, 1991). Attempts have been made to split the climate model grid square into more than one vegetation type. For example, Koster and Suarez (1992a,b), Verseghy *et al*. (1993), Ducoudre *et al*. (1993) and Bonan (1995) group the different types of vegetation within the grid square into 'clumps' or 'tiles' and perform separate calculations for each group (this is sometimes known as the grouped-mosaic approach). This allows for a differentiation between the major vegetation types and differs from the approach where the commonest vegetation type within the grid square is the only type modelled. A computationally expensive approach is to divide the climate model grid explicitly into a finer grid and simulate the land-surface–atmospheric exchanges for each of the finer grids (Seth *et al*., 1994). This is potentially very powerful if the division of the climate model grid extends into the boundary layer, since it aids the simulation of small-scale clouds.

A final area worthy of comment is the increasing significance of chemistry in land surface processes. Changes in CO_2 combine with changes in nitrogen deposition to affect many biogeochemical cycles. The biosphere may also play a significant role in the emission of cloud condensation nuclei and exert a strong influence on cloud radiative and microphysical properties (Andreae *et al*., 2002). These are areas that have not been included in LSM and represent a major new area of future research.

6. CONCLUSIONS

Land surface processes in climate models have evolved from a very simple, implicit approach representing the surface energy balance and hydrology (Manabe, 1969), to complex models that represent many of the key processes through which the land surface influences the climate simulated by climate models. The science has developed from individuals working independently, through to larger multidisciplinary teams. We now need expertise from the chemical, biological and geoscience communities to work with climate scientists to build LSM. We need the remote-sensing community, experts in geographical information systems and computer science to work with those in the biogeochemical and biophysical sciences. The infrastructure is being put in place whereby this multidisciplinary problem can be tackled by experts working from laboratories anywhere on Earth. New initiatives offer a genuine potential to realize a generation of balanced LSMs containing those components required to represent the role of the land surface in climate models. Success in this endeavour will enhance our capability to simulate the impacts of LCC and the impacts of increasing CO_2 on the global and regional environment.

The way forward for land surface modelling is thus to expand the climate modelling groups with the capacity to use third-generation LSMs and to improve the representation of hydrological processes in parallel. Although first- and second-generation LSMs could be developed, maintained and implemented into climate models by individuals or very small groups working independently, this is not the case with third-generation schemes, and moves to consolidate several LSMs into single, well-supported models have now begun (http://clm.gsfc.nasa.gov/). The breadth of knowledge required to remain at the cutting edge of the science is intimidating. Advanced understanding of soil temperature physics, soil moisture processes, large-scale hydrology, snow physics, radiative transfer, photosynthesis-level biochemistry and large-scale ecology, boundary-layer processes, biogeochemical cycling and advanced computer science are all required. This has to be coupled with a good understanding of climate and climate feedbacks. Thus, genuinely multidisciplinary teams are developing and collaborating to support these new models. Individuals are now thinking carefully about how to design these models for flexibility and to enable them to be linked into different climate models (Polcher *et al*., 1998). Broad international groups under the leadership of organizations like the World Climate Research Program (WCRP, http://www.wmo.ch/web/wcrp/wcrp/) and the International Geosphere Biosphere Program (IGBP, http://www.igbp.kva.se/) are forming or maturing. The Global Energy and Water Experiment (GEWEX, http://www.gewex.com/), which has driven the formation of the Global Land–Atmosphere System Study (GLASS, http://hydro.iis.u-tokyo.ac.jp/GLASS/), is one organizational group that recognizes that the complexity of the task is such that success requires genuine collaboration amongst a wide range of interested groups. GLASS includes the two major efforts over the last decade that have done most to bring the land surface community together to understand why various LSMs perform differently: PILPS (Henderson-Sellers *et al.*, 1995, 2002; http://www.cic.mq.edu.au/pilps-rice/) and the Global Soil Wetness Project (GSWP; Dirmeyer *et al.*, 1999; http://grads.iges.org/gswp/).

These initiatives will need to be successful. Even highly optimistic emission scenarios for greenhouse gases suggest that CO_2 will exceed 500 ppmv, nearly double pre-industrial levels, by 2100 (and may exceed 1200 ppmv, see Houghton *et al.* (2001)). Our vulnerability to resulting changes in climate is also increasing as population increases, especially in the provision of fresh water (e.g. see Vörösmarty *et al.*, 2000). Although modelling the full complexity of land surface processes and the interaction between these processes and climate is intimidating, remarkable progress has been made. Relatively new initiatives, whereby the land surface community has begun to liaise with the ecological and biogeochemical communities, provide us with a path forward to build integrated models of the terrestrial surface that should enhance our capability of modelling the Earth's climate into the 22nd century.

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