Atmosphere–terrestrial ecosystem interactions: implications for coupled modeling

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ABSTRACT

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This paper overviews the important interactions between ecosystem and atmospheric dynamics. Examples of suggested procedures to model these interrelationships are introduced. The concept of equilibrium versus non-equilibrium non-linear dynamics in the context of chaos is discussed. The paper concludes with a description of a parameterization of vegetation effects on atmospheric structure, including a model sensitivity simulation of the importance of land surface processes on this structure.

1. MODELING ATMOSPHERE–ECOSYSTEM INTERACTION

There are two approaches to atmospheric–terrestrial ecosystem interactive models: one-way interaction and two-way interaction. In a one-way interactive model, atmospheric or ecosystem characterization is specified as input to an ecosystem dynamic or meteorological model, respectively. An example of atmosphere-to-ecosystem linkage (Fig. 1) is to apply output from a general circulation model (GCM) through a mesoscale atmospheric model to an ecosystem model (Kittel and Coughenour, 1988). Such linkage permits an evaluation of regional meteorological effects on biological systems due to GCM-simulated changes, such as those due to greenhouse

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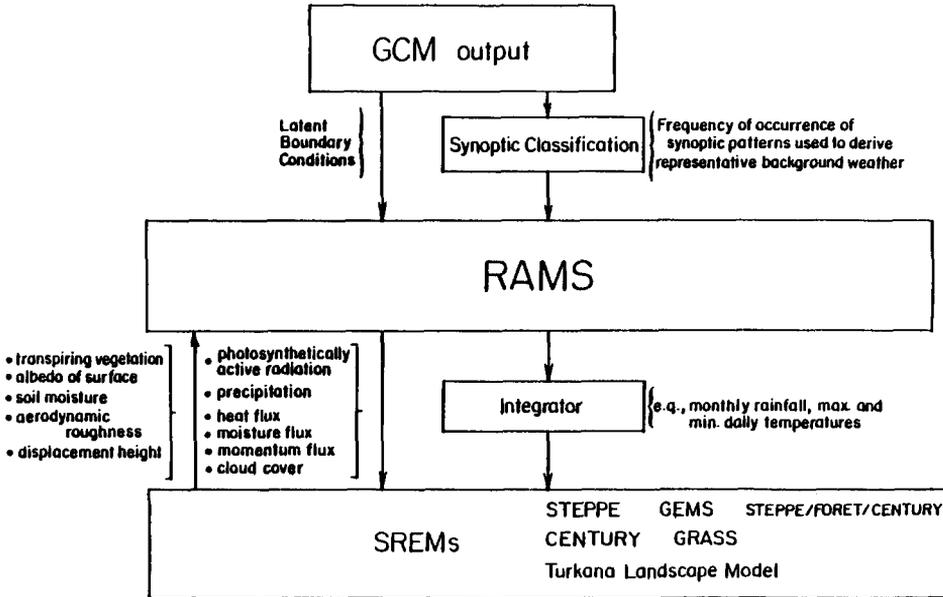
Fig. 1. Schematic of the procedure to interface general circulation models through a regional/mesoscale meteorological model (RAMS) to ecosystem dynamics models.

gas changes and anthropogenic aerosol input. This approach, using a mesoscale atmospheric model, provides spatially and temporally variable atmospheric impacts as inputs into an ecosystem model, which would not otherwise be available directly from the coarse-grid GCMs.

As an example of ecosystem-to-atmosphere one-way interaction, the effect of the land surface on climate is specified in regional or global climate models by land process models, such as those of Dickinson et al. (1986), Sellers et al. (1986) and Avissar and Pielke (1989). While these parameterizations of terrestrial processes permit vegetation to respond to short time period atmospheric changes (e.g., by alterations in time in stomatal conductance), there is no modeled procedure for vegetation structure to change over time with a resultant change in the subsequent mesoscale atmospheric weather. This requires a two-way interaction. Since the global climate consists of a spectrum of mesoscale climate such two-way interactions have global implications.

Simple response ecosystem models (SREMs) are constituted using physiologically based vegetation response models (Fig. 1) to simulate the seasonal development of vegetation and soil characteristics as well as changes in vegetation structure at annual and decadal scales, forced by atmospheric variables. It is essential to base SREMs on fundamental biophysical models

in order to develop a credible parameterized ecosystem code which can interface with the atmosphere. It is also necessary to permit the atmosphere to respond to these vegetation changes. This requires the development of two-way interactive atmosphere–ecosystem models through the coupling of climate and ecosystem models.

This paper briefly overviews in Section 2 the best available procedures to represent landscape variations in two-way coupled atmospheric/land surface models. Section 3 uses a simple energy budget to demonstrate the strong sensitivity of global climate to changes in land surface processes. Section 4 uses concepts in non-linear dynamics to illustrate that a two-way interactive atmosphere/land surface characteristics model may produce a chaotic response over time as contrasted to an equilibrium or periodic solution, if the interaction is sufficiently non-linear. Section 5 uses a one-way interactive one-dimensional model to illustrate the significant influence of land surface vegetation coverage on the overlying atmosphere.

2. LANDSCAPE VARIABILITY

Temporal and spatial variability of the land surface is not easily captured in equations analogous to the equations of motion in fluids. For coupled modeling, some of the spatial variability of the land surface must be described as boundary conditions, invariant or nearly so over the interval of model integration; these attributes include terrain and soil physical properties. The biotic characteristics of the land surface, however, do vary over time scales of seconds to decades, time scales relevant to modeling of the climate system. Much of the complexity of the land surface arises from features with a mix of time scales represented, ranging from landforms and bedrock reflecting history over geological time to stomates, which respond to changes in the atmosphere in seconds. All of these time scales are relevant to interactions of the atmosphere and land surface. Below, we review sources of spatial variability in the land surface relevant to the atmosphere, and how they may vary in time. We also discuss the best available techniques for including various sources of variability in coupled models of the atmosphere and land surface.

Topography and soil. Terrain and soils influence the atmosphere through physical interactions (i.e., roughness, flow separation), hydrology (water-holding capacity, runoff), and biology (effects of soil and topography on vegetation). For the purposes of most simulations, soils and terrain can be regarded as invariant, although investigations of the effects of orogeny on the atmosphere have shed significant light on the coupled evolution of the geosphere and climate system. Terrain is normally represented in models

using gridded digital elevation data (Dubayah et al., 1990). Soils are normally mapped as irregular polygons, reflecting their complex distribution in nature (Burke et al., 1990) but gridded global maps exist (reviewed in Stewart et al., 1989). Such maps describe the boundaries of soil types, as defined using standard taxonomic procedures. Soil variables required in coupled models (e.g., depth, hydraulic attributes, fertility) are not mapped directly and must be inferred from taxonomic descriptions. Soil mapping units may each contain significant variability in properties such as depth or texture, and so current soil data bases must be used judiciously in coupled modeling.

Vegetation structure. The interactions of the biota with atmospheric dynamics occur via vegetation structure (height, trees vs. grass or shrubs, density) and physiology (canopy conductance). Vegetation structure typically changes over periods of years to decades, reflecting biotic and biogeochemical processes such as succession (Pastor and Post, 1986), human disturbance, and episodic natural events such as wildfire or severe storms. Vegetation structure generally changes gradually, although stochastic extreme events can result in abrupt changes. Thus, vegetation structure tends to change at time scales associated with climate, rather than with "weather". Spatial variability in vegetation structure occurs due to terrain, soils, climate, and history (disturbance, land use) and can influence the dynamics of the atmosphere at regional scales (Avissar and Verstraete, 1990). Inclusion of vegetation dynamics over time scales of years to decades in climate system models is one of the most serious challenges to the development of robust coupled models in the coming years.

Plant physiology. Physiological responses also vary in time and space, in a complex fashion. While leaf-level resistance to water vapor changes rapidly in response to drought in the atmosphere (seconds–hours), canopy resistance, the integral of leaf resistance over the canopy plus an aerodynamic factor, changes at several time scales (Sato et al., 1989). At one scale is the rapid plant response to radiation and atmospheric moisture that change diurnally and move rapidly in response to cloud cover changes, atmospheric boundary layer growth, and advection of air masses. Next is the slower response of stomatal resistance to soil moisture, which varies over days to a few weeks. In the longer term, canopy resistance is regulated by leaf area. Studies in both semiarid forests (Running and Nemani, 1989) and grasslands showed that leaf area tends to equilibrate with soil moisture availability to homogenize leaf moisture stress. This equilibrium of leaf area with water availability results in a seasonal component in the dynamics of canopy resistance.

Nutrient limitation may also constrain leaf area development, especially in moist environments (Schimel et al., 1991a). This can add additional levels of complexity to the long-term responses of canopy resistance to changing conditions (Schimel et al., 1990, 1991a). Nutrients also vary in space with soil attributes and vegetation development and so contribute to the spatial variability of leaf area (Pastor and Post, 1986; Schimel et al., 1990).

Recent studies conducted as part of FIFE (First ISLSCP Field Experiment) demonstrated complex interactions between landscape variability in water storage (soil depth), management practices affecting evaporation rate, and the response of leaf and canopy physiology (Schimel et al., 1991a; Davis et al., 1992; Turner et al., 1992). These studies demonstrated that complex spatial and temporal patterns in canopy physiology could be explained by simple models based on a few principles (Schimel et al., 1991b). This result is encouraging for the future development of coupled models.

Implications for coupled model development. While simple simulations of physiological coupling of vegetation to the atmosphere have been developed and applied successfully, it is clear that two problems now motivate the development of a new class of models. First, simulation of vegetation changes over the time scales at which leaf area changes and longer requires the simulation of plant growth and allocation in order to model lower frequency changes in canopy resistance. Running et al.'s (1989) Forest BGC model is an example of such a model. Second, these models must be coupled to spatial data bases describing topographic and edaphic features, in order to adequately describe spatial variability in surface processes (Burke et al., 1990, 1991).

Long-term simulations show that interactions of plant processes with soil nutrient turnover influence the response of vegetation to changing or variable climate (Schimel et al., 1990; Pastor and Post, 1988). Thus, models of land surface–atmosphere coupling must also include biogeochemical interactions in order to represent dynamics over periods of years to decades. While there will be exciting developments and fundamental discoveries in the behavior of the coupled land surface–atmosphere system (Watson and Lovelock, 1983; Zeng et al., 1990), ongoing studies have already revealed rich and interesting behavior.

3. CLIMATE SENSITIVITY TO LAND SURFACE CHANGE

It is clear from a simplified analysis that there is a significant sensitivity of climate and weather to land surface processes in even small changes in

vegetation structure. Pielke and Avissar (1990) demonstrated the sensitivity of the Earth's equilibrium temperature to small changes in land surface albedo. Here we more precisely derive this analysis. Long wave radiation emitted from over the entire Earth's surface balances solar radiation impinging on a disc with the Earth's diameter. The balance is expressed as:

$$\sigma T_E^4 = \tfrac{1}{4}(1 - A)S \tag{1}$$

where σ is the Stefan–Boltzmann constant (5.67×10^{-8} W m^{-2} K^{-4}), T_E is the Earth's equilibrium temperature, A is the Earth's planetary albedo, and S is the solar flux at the top of the atmosphere. This equation is a more accurate representation of the bulk of Earth's radiative energy balance than given in the Pielke and Avissar article. Differentiation of eq. (1) with respect to albedo and assuming small changes, gives:

$$\Delta T_E = -1.52 \times 10^9 (\text{K}^4)\frac{\Delta A}{T_E^3}$$

where $S = 1380$ W m^{-2} has been specified. For different values of equilibrium temperature, T_E:

$$
\begin{aligned}
\Delta T_E &\cong -83\Delta A, \quad \text{for } T_E = 263 \text{ K} \\
\Delta T_E &\cong -67\Delta A, \quad \text{for } T_E = 283 \text{ K} \\
\Delta T_E &\cong -63\Delta A, \quad \text{for } T_E = 288 \text{ K} \\
\Delta T_E &\cong -54\Delta A, \quad \text{for } T_E = 303 \text{ K}
\end{aligned} \tag{2}
$$

Because the land surface only covers about one-quarter of the planet, eq. (2) for $T_E = 283$ K, for example, becomes:

$$\Delta T_E = -17\Delta A_{\text{land}}$$

such that a 10% change in land surface albedo (A_{land}) as viewed from outside the Earth's atmosphere results in a 1.7°C equilibrium temperature change. For $T_E = 263$ K a 2.1°C change results. Thus, a 4% increase in land surface albedo would result in a 0.7 to 0.8°C decrease in equilibrium temperature, which is of the same magnitude as the current expected effect of greenhouse gas warming. Indeed, it is not surprising that such a small effect is important because a similar sensitivity has been found for clouds.

The sensitivity of global equilibrium temperature to land albedo change may additionally be influenced by the latitudinal location of albedo changes (predominantly in the lower and middle latitudes), the relative contribution to the planetary albedo in these regions of clouds versus the land, and positive or negative feedback between surface albedo and cloud cover (e.g., Charney et al., 1977). An analysis of the sensitivity of global surface temperature rather than that of equilibrium temperature, of course, is far

more complex because of interactions between radiation, clouds, atmospheric dynamics, and surface biologic and hydrologic processes. Nonetheless, this simple analysis demonstrates the strong sensitivity of global climate to surface characteristics.

4. EQUILIBRIUM VS. NON-EQUILIBRIUM DYNAMICS IN THE CONTEXT OF CHAOS

As mentioned in the previous sections, there exist strong interactions between the atmosphere and ecosystems. These interactions may lead to equilibrium state (steady state) or non-equilibrium state (oscillation or chaos) behavior. Equilibrium state means that the final state of a system approaches a fixed value after transients in the system die away. Oscillation means that the final state of a system oscillates among more than one fixed value. Chaos means that the final state oscillates among an infinite number of values and that small differences in initial conditions grow exponentially with time and lead to large variance in final states or even lead to completely different final states (i.e., sensitivity to initial conditions). In addition, a chaotic system has a fractal rather than integer dimension. Sensitivity to initial conditions is quantitatively expressed by positive Lyapunov exponents which can be computed by various methods (Zeng et al., 1991). Fractal dimension can be obtained by different algorithms (Zeng et al., 1992). In a system's power spectra (Fourier transformation of a time series in the system), an equilibrium state is characterized by a straight line, oscillation by one or more peaks, and chaos by the presence of broadband noise (Bergé et al., 1984).

The term "chaos" is reserved by many investigators for irregularity in deterministic dissipative dynamic systems. Deterministic systems are usually described by partial or ordinary differential equations, difference equations, or simple iterative algebraic equations, where all coefficients contain no random perturbations. Given accurate initial conditions, only one deterministic solution will be obtained. However, there are always some errors in initial conditions (e.g., measurement errors or computer truncation errors), and these small errors grow exponentially with time. This error growth is solely determined by non-linear dynamics in a chaotic system. In contrast, randomness or a lack of complete determinism always occurs in stochastic systems because of a stochastic component of the governing equations. The definition of chaos may be broadened to include stochastic systems, provided that it appears that the stochastic system would remain aperiodic if the stochastic part of the governing equations was removed. This modification makes it possible to include real physical systems, whose actual determinism is likely to be in doubt (Lorenz, 1987).

Next, we illustrate these concepts by means of the logistic difference equation which is arguably the simplest non-linear difference equation. This equation may explain population growth in population ecology theory and many other phenomena in nature. It is written as:

$$X_{n+1} = \mu X_n (1 - X_n)$$

where the subscript n is a time step index $(0, 1, 2, \ldots)$, X is the state variable and ranges between 0 and 1, and the parameter μ is between 0 and 4.

For $0 < \mu < 1$, all trajectories are attracted to the equilibrium state $X = 0$. For $1 < \mu < 3$, we still obtain an equilibrium state, i.e. $X = 1 - 1/\mu$. Although the final state approaches an equilibrium as $n \to \infty$, there may be some initial damping oscillations. For $\mu_1 = 3 \le \mu \le 3.449 = \mu_2$, X_n oscillates between two values. The process by which the final state changes from one equilibrium state to two equilibrium states as μ increases is called bifurcation, and μ_1 is called the bifurcation point. For $\mu_2 = 3.449 \le \mu \le 3.545 = \mu_3$, X_n oscillates among four equilibrium states, and μ_2 is the bifurcation point from period 2 to period 4. In this way, periodic bifurcation occurs as μ is further increased. When μ is greater than $\mu_\infty \approx 3.567$, chaos occurs. As μ is increased from μ_1 to μ_∞, the difference between bifurcation points becomes smaller and smaller.

We generalized the concept of chaos to an idealized ecosystem/ atmospheric interactive model called Daisyworld. Daisyworld is a cloudless flat or cylindrical planet with negligible atmospheric greenhouse gases that bears life only in the form of two different species of daisies. The Daisyworld model was originally presented by Watson and Lovelock (1983) as an idealized example of the Gaia hypothesis. In this hypothesis, the environment restricts the evolution of the biota and dictates what type of life can exist as a consequence of Darwinian natural selection, and the biota regulate and stabilize automatically the environment in response to alterations in external forcing such as solar luminosity. In the Daisyworld model, daisies with different albedos evolve in land coverage in response to changes in solar luminosity, with the daisies with greater albedo preferentially developing when the sun's output increases. Zeng et al. (1990) showed that an equilibrium state can be obtained when the coupling between daisies and environmental temperature is weak. When the nonlinear interaction between them is sufficiently strong, however, periodic and chaotic responses occur. Time series of areal coverage of black and white daisies and the power spectrum of the model demonstrate the chaotic response (Fig. 2). The chaotic response was also verified mathematically by evaluating the fractal dimension and Lyapunov exponents (Zeng et al., 1990). The implication for the actual atmosphere–ecosystem interaction
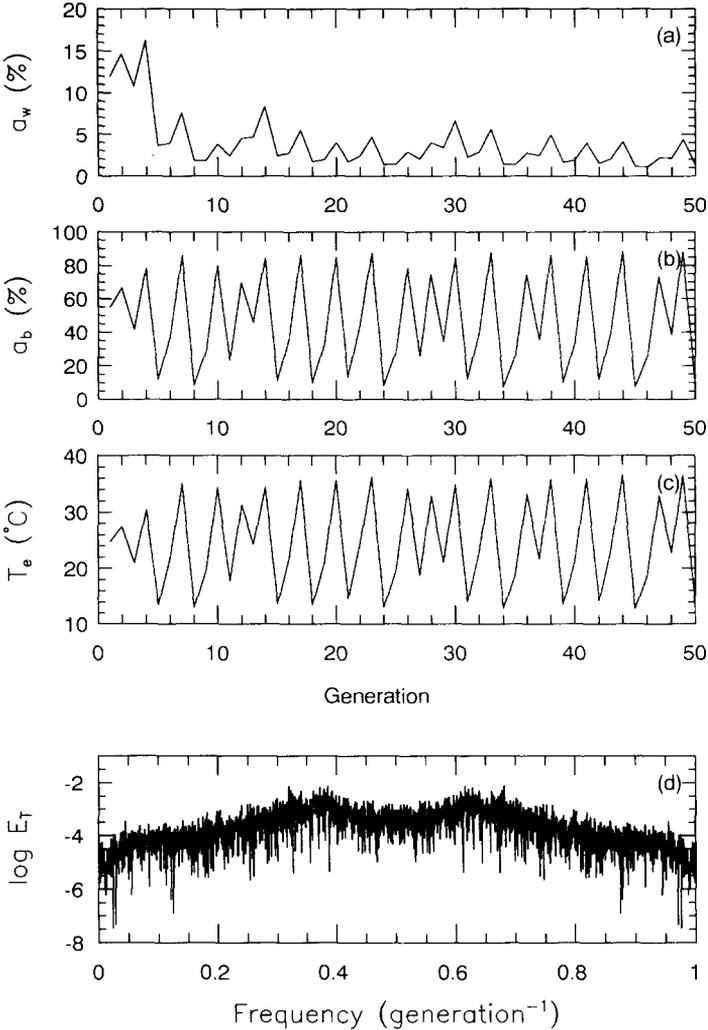
Fig. 2. Variation with generation of (a) percent of land area of white daisies; (b) percent of land area of black daisies; (c) surface temperature; and (d) spectrum of temperature variation as a function of frequency from the Daisyworld model (Zeng et al., 1990).

is that man-caused landscape changes could result in very substantial irregular changes in the climate.

5. SHORT-TERM INTERACTIONS

The biospheric models existing in atmospheric modeling systems are currently steady state models. Parameters commonly used in ecosystem dynamics (e.g., LAI, biomass) are specified at a given time in a year and do

not change from year to year. Only with the recent work by Henderson-Sellers (1990), have there been preliminary attempts to construct a two-way interactive numerical model. With respect to short-term interactions, it has been demonstrated that changes of land use patterns can alter the structure of the atmospheric boundary layer and hence, the associated convective clouds (Segal et al., 1988, 1989; Rabin et al., 1990). Since the time scale of change of ecosystem structure is much longer (usually on the order of seasons to decades) than the time scale of change of atmospheric boundary layer structure (on the order of hours), one-way interaction has been used exclusively in recent studies of the atmospheric response to changes in land use. In this case the ecosystem is assumed to be steady and the atmosphere
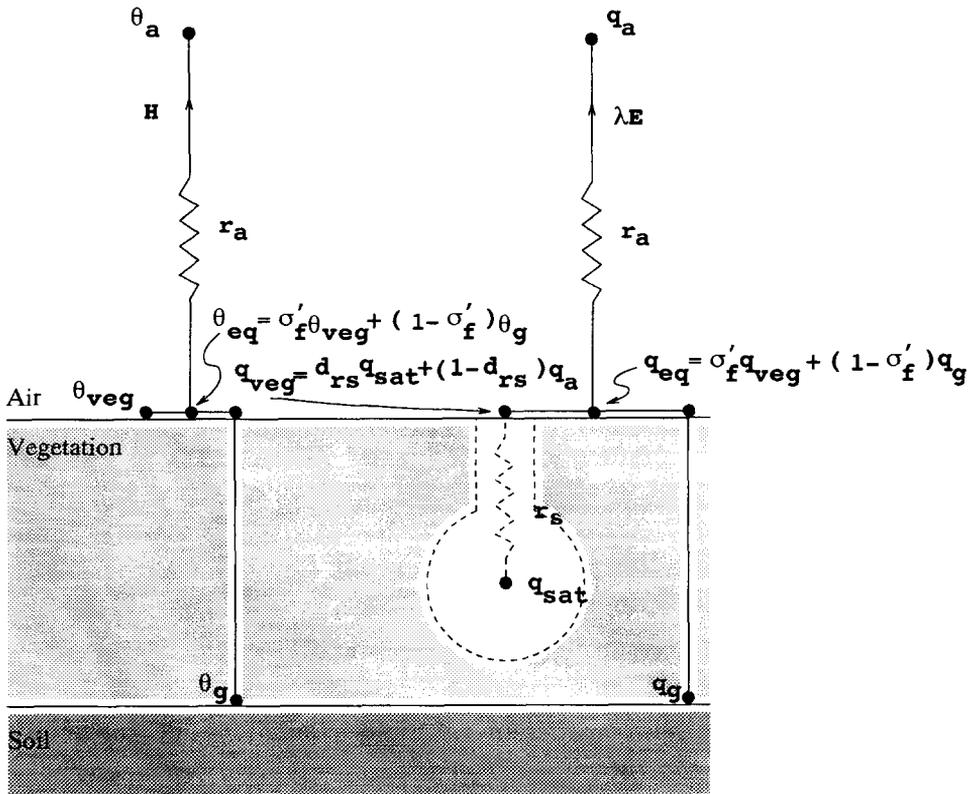

Fig. 3. Resistance diagram of the soil–vegetation–atmosphere system. Variable θ is potential temperature, q is specific humidity, H is sensible heat flux, E is latent heat flux, h is aerodynamic resistance and d_{rs} is the relative stomatal conductance. Subscript a indicates the variable for the air, veg indicates the variable for the canopy, and g indicates the variable for the ground. Variable σ_f' is an effective ground coverage function as described by Avissar and Pielke (1989).

responds to the change in the biophysical state (stressed or unstressed vegetation). Future work is needed, however, to determine the extent to which changes in boundary layer structure and associated clouds and precipitation feedback to the vegetation characteristics. The model discussed below could be linked to an ecosystem dynamics model to accomplish this goal.

The Land Ecosystem–Atmosphere Feedback (LEAF) model used in the Colorado State University (CSU) Regional Atmospheric Modeling System (RAMS) is characterized as a canopy–soil "big-leaf" model (Goudriaan, 1989). Incoming radiation and weather conditions can be predicted by an atmospheric model or specified as a boundary condition. The amount of leaf area and plant biomass are prescribed as model parameters. The canopy conductance is defined as the total LAI times the stomatal conductance, which is a function of radiation, soil water potential, vapor pressure deficit and leaf temperature (Avissar and Pielke, 1989). The underlying soil is parameterized according to McCumber and Pielke (1981) and Lee and Pielke (1992). Figure 3 shows that the resistance function has been absorbed into a relative stomatal conductance function in the LEAF model. Figure 4 shows the model predicted partitioning of energy on an early
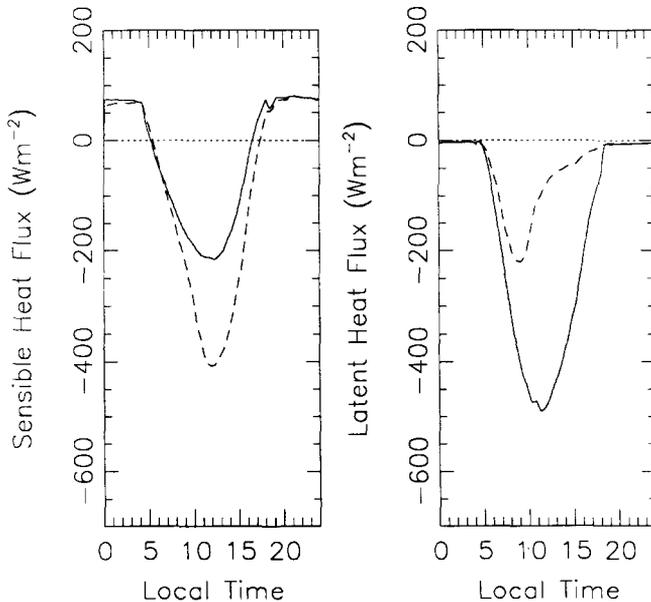

Fig. 4. Sensible heat flux (left panel) and latent heat flux (right panel) for vegetated (solid lines) and bare soil (dashed lines) surface for June conditions in Kansas.

summer day (June 6) in eastern Kansas into sensible and latent heat fluxes with and without the presence of vegetation. When vegetation (tall grass in this case) is present, most of the energy goes into the latent heat flux and the evapotranspiration is large. When the surface is bare (silty clay loam soil), the incoming net radiation first dries out the top soil layer and create a local maximum of latent heat flux returned to the atmosphere. However, most of the incoming net radiation goes into sensible heat flux returned to the atmosphere later in the day. It is well understood that the atmospheric boundary layer should be deeper when the sensible heat flux is large and shallower when the sensible heat flux is small. On the other hand, convective clouds are formed more easily down-wind of lakes and heavily irrigated areas as documented in Rabin et al. (1990). This example shows a typical short-term interaction between the ecosystem and the atmosphere. A next step using this model is to ascertain species composition changes that would occur due to the different overlying mesoscale meteorological conditions.

6. SUMMARY

This paper overviews the important interactions between ecosystems and atmospheric dynamics. Examples of suggested procedures to model these interrelationships are introduced. The concept of equilibrium versus non-equilibrium non-linear dynamics in the context of chaos is discussed. The paper concludes with a description of a parameterization of vegetation effects on atmospheric structure, including a model sensitivity simulation of the importance of land surface processes on this structure.

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