ABSTRACT

This paper reviews the cumulative effect of land-cover change on local, regional, and global weather and climate. Documented land-cover changes from 1700 to the present are shown to be particularly significant in the tropics. Model results are presented which show the teleconnection of weather patterns worldwide result from these landscape changes.

1. Introduction

Pielke (1984; 2002) defines mesoscale atmospheric features as those in which the horizontal winds are significantly different from the gradient wind, even above the planetary boundary layer, but where the pressure distribution is accurately represented by the hydrostatic equation. The validity of the hydrostatic equation permits a diagnosis of the pressure field at any time from the spatial distribution of the temperature field. The failure of the gradient wind relationship to represent the nonturbulent wind field means that the wind field cannot be diagnosed from the pressure field at the same time. When the gradient wind relation is valid, the instantaneous pressure field can be used to accurately represent the actual wind.

Lynch (1988) demonstrates that the horizontal wind can be represented by

$$\vec{V} = \vec{V}_R + \vec{V}_D$$

where $\vec{V}_R$ is the rotational component of the wind and $\vec{V}_D$ is the divergent component.
When the winds are in gradient wind balance, $|\vec{V}_r| \gg |\vec{V}_D|$, since

$$\vec{V}_r = \vec{V}_r \approx \frac{k}{f} \nabla_p \phi$$

where $g$ is the gravitational acceleration, $p$ is pressure, $f$ the Coriolis parameter, and $\phi$ is the geopotential height ($\phi = g z$). Such an approximation works well in the middle and high latitudes, where $f$ is larger. However, it fails in the tropics. As a result $|\vec{V}_r|$ and $|\vec{V}_D|$ are often of the same magnitude even for larger atmospheric features.

Thus, mesoscale systems are those in which $|\vec{V}_D|$ is a significant fraction of $|\vec{V}_r|$. In the tropics (where $f$ is small), all weather features are, therefore, mesoscale, even though their horizontal extent can be across thousands of kilometers in the longitudinal direction. In this chapter, however, since larger tropical systems (e.g., monsoons) are covered elsewhere, the focus is on reported scales of several hundreds of kilometers and less.

Another distinguishing feature of tropical systems is that the vertical gradient of the equivalent potential temperature, $\theta_e$, is frequently less than zero. This means that when saturation is realized, cumuliform clouds will occur. Cumulonimbus clouds, therefore, are a characteristic of the tropical atmosphere and of mesoscale systems. The excursion of tropical air masses into the middle, and even high latitudes, in the warmer seasons, is why thunderstorms often occur at that time of the year, while stratiform-precipitating cloud systems are typical in the cooler seasons. Thunderstorms are also the reason that tropical land-use change and dynamics can rapidly teleconnect globally.
2. Land-Use History of the Tropics

Land areas of the tropics have undergone increasing intensive alterations since 1700. Figure 1 documents the best estimate of this change. Large areas of Asia were already disturbed by human activities prior to 1700. Of large tropical regions, only the Amazon region has been relatively undisturbed until recently. Figure 2 illustrates the vegetation changes that exist because of human disturbance.

Table 1 documents the changes during the last several decades of tropical land areas. Clearly evident from this table is the continuing conversion of tropical forest to agricultural and grazing lands.

An important research question with very practical societal consequences is the effect of this conversion on local, regional, and global climate. This subject is discussed in the next section of this chapter.

3. Effect of Landscape on Local Climate

3.1 Surface Effects

Pielke (2001) reviews the role of vegetation and soils on the prediction of cumulus convective rainfall. That paper is used as the basis to describe how land use and its change over time alter the local climate. Text from that paper is adapted for the following summary.

The surface energy and moisture budgets for bare and vegetated soils are schematically illustrated in Figures 3 and 4. These surface budgets can be written as

\[ R_N = Q_G + H + L(E+T) \]  

\[ P = E + T + RO + I \]
where $R_N$ represents the net radiative fluxes $= Q_s (1 - A) + Q_{\text{LW} \downarrow} - Q_{\text{LW} \uparrow}^\iota; P$ is the precipitation; $E$ is evaporation (this term represents the conversion of liquid water into water vapor by non-biophysical processes, such as from the soil surface and from leaves and branches); $T$ is transpiration (represents the phase conversion to water vapor, by biological processes, through stoma on plants); $Q_G$ is the soil heat flux; $H$ is the turbulent sensible heat flux; $L (E + T)$ is the turbulent latent heat flux; $L$ is the latent heat of vaporization; $RO$ is runoff; $I$ is infiltration; $Q_s$ is solar insolation; $A$ is albedo; $Q_{\text{LW} \downarrow}^\iota$ is downwelling longwave radiation; $Q_{\text{LW} \uparrow}^\iota$ is upwelling longwave radiation $= (1 - \varepsilon) Q_{\text{LW} \downarrow}^\iota + \varepsilon \sigma T_s^4; \varepsilon$ is the surface emissivity; and $T_s$ is the surface temperature.

Detailed discussion of these terms is given in Pielke (2002; Chapter 11). Equations (1) and (2) are not independent of each other. A reduction in evaporation and transpiration in Eq. (2), for example, increases $Q_G$ and/or $H$ in (1) when $R_N$ does not change. This reduction can occur, for example, if runoff is increased (such as through clear-cutting a forest). The precipitation rate (and type) also obviously influences how water is distributed between runoff, infiltration, and the interception of water on plant surfaces.

The relative amounts of turbulent sensible ($H$) and latent heat fluxes [$L(E + T)$] are used to define the quantity called the Bowen ratio ($B$), and evaporative fraction, $e_f$;

$$B = \frac{H}{L(E + T)}; \quad e_f = L(E + T)/R_N. \quad (3)$$

The denominator $L(E + T)$ has been called “evapotranspiration,” although since evaporation and transpiration involve two distinct pathways for liquid water to convert to water vapor, the use of the term “evapotranspiration” should be discouraged. The relation of $R_N$ to $H$ and $L(E + T)$, following Segal et al. (1988), can be written as
With $Q_G \ll H$ and $E + T$, as discussed in Segal et al. (1988),

$$H \approx \frac{R_N Q_G}{(1/B) + 1}.$$ 

Segal et al. (1995) showed that with the same value of $R_N$, with a smaller Bowen ratio, the thermodynamic potential for deep cumulus convection increases.

Therefore, any land-use change that alters one or more of the variables in Eqs. (1) and (2) will directly affect the local climate. For instance, a decrease in $A$ (i.e., a darkening of the surface) would increase $R_N$; thus making more heat energy available for $Q_G$, $H$, $E$, and $T$. The heat that goes into $H$ increases $\theta$ since $T$ increases. The heat that goes into $E$ or $T$ goes into $\theta_e$ since $w$ increases. If the surface were dry and bare, all of the heat energy would necessarily go into $Q_G$ and $H$ as shown by Pielke (1984; page 381) for the empty quarter in Saudi Arabia.

Lyons et al. (1996), for example, found a reduction of $H$ in southwestern Australia as a result of the conversion of land to agriculture. Bryant et al. (1990) found higher sensible heat fluxes in the Sonoran Desert of Mexico due to overgrazing. Fitzjarrald et al. (2001), Freedman et al. (2001), and Schwartz (1994) found that the leafing out of vegetation in the spring has a dramatic effect on a reduction in $H$. Schriberger et al. (1996) and Rabin et al. (1990) discuss how cumulus cloud base height is directly related to surface heat and moisture fluxes, as modulated by the characteristics of the underlying heterogeneous surface. Trace gas fluxes and biogenic aerosol emissions and deposition will, of course, also be altered as landscape changes (see Pielke 2002; Chapter 7, as to how these effects are coupled to Eq. 1).

A conclusion of this analysis is that changes in the Earth’s surface can result in significant changes in the surface energy and moisture budgets.
3.2 Boundary-Layer Effects

Once the surface energy budget is altered, fluxes of heat, moisture, and momentum within the planetary boundary layer are directly affected (Segal et al. 1989). As an example, Fig. 5 illustrates an idealization of the vertical structure of the convective boundary layer, where the surface heat flux, $H$, depth of the layer $z_i$, and the temperature stratification just above determine the vertical profile of temperature and heat flux. Deardorff (1974) suggested a growth rate equation for $z_i$ in the absence of large-scale wind flow which is proportional to

$$\frac{\partial z_i}{\partial t} \approx H^{2/3} z_i^{4/3}. \quad (4a)$$

The entrainment of air from above $z_i$ to heights below $z_i$ is given by

$$H_{z_i} = \alpha H \quad (4b)$$

where $\alpha$ is the entrainment coefficient ($\alpha \approx 0.2$, although there are suggestions it is different from this value; Betts et al. 1992). McNider and Kopp (1990) discuss how the size of thermals generated from surface heating are a function of $z_i$, $H$, and height within the boundary layer. The rate of growth of the boundary layer during the day, and the ingestion of free atmospheric air into the boundary layer are, therefore, both dependent on the surface heat flux, $H$.

A simplified form of the prognostic equation for $\theta$ can be used to illustrate how temperature change is related to the surface heat flux, $H_s$,

$$\frac{\partial \theta}{\partial t} = \frac{\partial}{\partial z} \left( \frac{H}{\rho C_p} \right),$$

where $\rho$ is the air density, and $C_p$ is the specific heat at constant pressure. Integrating from the surface to $z_i$, and using the mean value theorem of calculus yields
\[ \frac{\partial \theta}{\partial t} = \frac{1}{z_i \rho_c} \left[ H_z - H_{z_i} \right] = \frac{1.2}{\rho_c z_i} H_z \]

where Eq. (4b) with \( \alpha = 1.2 \) has been used. Using this equation, a heating rate of a 1 km deep boundary layer of 2°C over 6 hours is produced by a surface heat flux of 100 W m\(^{-2}\).

A conclusion from this section and associated references, is that the boundary-layer structure, including its depth, are directly influenced by the surface heat and moisture fluxes.

### 3.3 Local Wind Circulations

Local wind circulations can subsequently result from horizontal variations in \( H \) and \( z_i \) (Segal and Arritt 1992). Such wind circulations are referred to as “solenoidal circulations” and are the reason sea- and land-breezes occur (Simpson 1994; Pielke 1984, Chapter 13). The reason that these local wind circulations can develop due to these variations is discussed in Pielke and Segal (1986).

Such mesoscale circulations can produce focused regions of particularly favorable conditions for thunderstorms (Pielke et al. 1991a). In these areas, CAPE, and other measures of the potential for deep cumulus convection are increased in response to boundary wind convergence associated with the local wind circulations (Pielke et al. 1991b). Convective inhibition is reduced in these areas. These wind convergence zones can also provide specific vertical motion “triggers” with which to initiate deep cumulus convection. Therefore, the spatial structure of the surface heating, as influenced by landscape, can produce focused regions for deep cumulonimbus convection.

### 4. Mesoscale and Regional Horizontal Perspective

Since different landscape characteristics result in varying values of boundary layer structure, mesoscale atmospheric circulations can result from the landscape heterogeneity as discussed in previously. Studies of this mesoscale effect include Ookouchi et al. (1984), Mahfouf et al.
(1987), Segal et al. (1988), Pielke et al. (1991a), Segal and Arritt (1992), Avissar and Chen (1993), Chen and Avissar (1994a), Goutorbe et al. (1994), Mahrt et al. (1994), Lynn et al. (1995a), Avissar and Liu (1996), Stohlgren et al. (1998), Taylor et al. (1998), Wang et al. (1997, 1998), and Chase et al. (1999). At least one study, however, has found no significant effect, in general, due to real-world landscape heterogeneities (Zhong and Doran 1998, Doran and Zhong 2000). The conclusion of this latter work is disagreed with in Weaver and Avissar (2001), where they document observationally and using the same model as applied by Zhong and Doran that landscape heterogeneity in Oklahoma and Kansas does produce organized areas of cumulus convection.

Since land-water contrasts permit the development of sea breezes, which focus thunderstorm development over islands and coastal regions in the humid tropics, and in humid middle and high latitudes during the summer (e.g., see Fig. 12-13 from Pielke 1984), it would be expected that similar variations in surface heating associated with landscape patterns would also produce mesoscale circulations of a similar magnitude. Other papers which describe the influence of sea breezes on thunderstorms are listed in Pielke (1974) and Pielke et al. (1991b).

Avissar and Schmidt (1998) have explored how landscape patchiness influences cumulus development using a large eddy simulation. They found preferential locations within the heterogeneous landscape where pockets of relatively high moisture concentrations occurred. As shown in Figures 6 and 7, the shape of the heterogeneity strongly influences the ability of mesoscale flows to concentrate CAPE within local regions so as to permit a greater likelihood of stronger thunderstorms. The large square shaped area, for example, is able to focus the lower tropospheric winds so as to optimize the accumulation of CAPE. This focusing of CAPE is analogous to what occurs with round islands (Neumann and Mahrer 1974). Dalu et al. (1996)
used a linear model to conclude that the Rossby radius defined as Eq. (A41) in Pielke (2001) is
the optimal spatial scale for landscape heterogeneities to produce mesoscale flows. Avissar and
et al. (1997), and Avissar and Schmidt (1998), also explored the issue of the size of landscape
patchiness that is needed before the boundary-layer structure is significantly affected and a
mesoscale circulation produced. Consistent with these conclusions, Segal et al. (1997) found
that cumulus clouds are a minimum downwind of mesoscale-sized lakes during the warm season
as a result of mesoscale-induced subsidence over the lake and the resultant suppression of $z_l$.

Other studies which have explored the influence of landscape heterogeneity on cumulus
convection include Segal et al. (1989), Rabin et al. (1990), Chang and Wetzel (1991), Fast and
McCormicle (1991), Segal and Arritt (1992), Chen and Avissar (1994a,b), Li and Avissar (1994),
Clark and Arritt (1995), Cutrim et al. (1995), Lynn et al. (1995a,b, 1998), Rabin and Martin
(1996), and Wang et al. (2000).

Clark and Arritt (1995) found while the cumulus cloud precipitation was delayed when
the soil moisture was higher, the total accumulation of precipitation was greater. The largest
rainfall was generally predicted to occur for moist, fully vegetated surfaces. De Ridder and
Gallée (1998) found significant increases in convective rainfall in southern Israel associated with
irrigation and intensification of agricultural practices, while De Ridder (1998) found that dense
vegetation produces a positive feedback to precipitation. Baker et al. (2001) explored the
influence of soil moisture and other effects on sea-breeze initiated precipitation in Florida.

Emori (1998) shows, using idealized simulations, how cumulus rainfall and soil moisture
gradients interact so as to maintain a heterogeneous distribution of soil moisture. Taylor et al.
(1997) concluded that such a feedback occurs in the Sahel of Africa which acts to organize
cumulus rainfall on scales of about 10 km. Simpson et al. (1980, 1993) have shown that cumulus
clouds that merge together into a larger scale produce much more rainfall.

The effect of landscape evaporation and transpiration on thunderstorms is quite nonlinear.
These opposing effects further explain the apparent contradiction between the results reported in
Lyons et al. (1996) and Pielke et al. (1997). While increased moisture flux into the atmosphere
can increase CAPE, the triggering of these deep cumulus clouds may be more difficult since the
sensible heat flux may be reduced. The depth of the planetary boundary layer, for example, will
be shallower, if the sensible heat flux is less. Other examples of studies which explore how
vegetation variations organize cumulus convection include Anthes (1984), Vidale et al. (1997),
Liu et al. (1999), Souza et al. (2000), and Weaver et al. (2000).

There are also studies of the regional importance of spatial and temporal variations in soil
moisture and vegetation coverage (e.g., Fennessy and Shukla 1999; Pielke et al. 1999a).
Delworth and Manabe (1989) discuss how soil wetness influences the atmosphere by altering the
partitioning of energy flux into sensible and latent heat components. They found that a soil
moisture anomaly persists for seasonal and interannual time scales so that anomalous fluxes of
sensible and latent heat also persists for long time periods. A similar conclusion was reported in
Pielke et al. (1999a). Wei and Fu (1998) found that the conversion of grassland into a desert in
northern China would reduce precipitation as a result of the reduction in evaporation. Jones et al.
(1998) discuss how surface heating rates over regional areas are dependent on surface soil
wetness. Viterbo and Betts (1999) demonstrated significant improvement in large-scale
numerical weather prediction when improved soil moisture analyses were used. Betts et al.
(1996) reviewed these types of land-atmosphere interactions, as related to global modeling.
Nicholson (2000) reviewed land-surface processes and the climate of the Sahel. Other recent
regional-scale studies of the role of landscape processes in cumulus convection and other aspects of weather include Lyons et al. (1993), Carleton et al. (1994), Copeland et al. (1996), Huang et al. (1996), Bonan (1997), Sun et al. (1997), Bosilovich and Sun (1999), Liu and Avissar (1999a,b), Adegoke (2000), Li et al. (2000), Mohanty et al. (2001), and Niyogi et al. (2002a). A specific summary for India is reported in Niyogi et al. (2002b).

Kanae et al. (1994) concluded that deforestation in southeastern Asia since 1951 has resulted in decreases in rainfall in September in this region, when the large-scale monsoon flow weakens. Kiang and Eltahir (1999), Eastman et al. (2001), Lu et al. (2001), and Wang and Eltahir (1999, 2000a,b) have used coupled regional atmospheric-vegetation dynamics models to demonstrate the importance of two-way interaction between the atmosphere and vegetation response. Hoffman and Jackson (2000), for example, propose that as a result of atmospheric-vegetation interactions in tropical savanna regions, anthropogenic impacts can exacerbate declines in precipitation. Shinoda and Gamo (2000) used observations to demonstrate a correlation between vegetation and convective boundary-layer temperature over the African Sahel.

5. Global Perspective

The effect of well-above average ocean temperatures in the eastern and central Pacific Ocean, which is referred to as “El Niño”, has been shown to have a major affect on weather, thousands of kilometers from this region (Shabbar et al. 1997). The presence of the warm ocean surface conditions permits thunderstorms to occur there that would not have happened with the average colder ocean surface. These thunderstorms export vast amounts of heat, moisture, and kinetic energy to the middle and higher latitudes, particularly in the Winter Hemisphere. This
transfer alters the ridge and trough pattern associated with the polar jet stream (Hou 1998). This transfer of heat, moisture, and kinetic energy is referred to as “teleconnections” (Namias 1978, Wallace and Gutzler 1981, Glantz et al. 1991). Almost two-thirds of the global precipitation occurs associated with mesoscale cumulonimbus and stratiform cloud systems located equatorward of 30° (Keenan et al. 1994). In addition, much of the world's lightning occurs over tropical land masses, with maximums also over the midlatitude land masses in the warm seasons (Lyons 1999; Rosenfeld 2000). These tropical regions are also undergoing rapid landscape change (O’Brien 2000).

As shown in the pioneering study by Riehl and Malkus (1958), and Riehl and Simpson (1979), 1500 to 5000 thunderstorms (which they refer to as “hot towers”) are the conduit to transport this heat, moisture, and wind energy to higher latitudes. Since thunderstorms only occur in a relatively small percentage of the area of the tropics, a change in their spatial patterns would be expected to have global consequences.

Wu and Newell (1998) concluded that sea surface temperature variations in the tropical eastern Pacific Ocean have three unique properties that allow this region to influence the atmosphere effectively: large magnitude, long persistence, and spatial coherence. Since land-use change has the same three attributes, a similar teleconnection would be expected with respect to landscape patterns in the tropics. Dirmeyer and Shukla (1996), for example, found that doubling the size of deserts in a GCM model caused alterations in the polar jet stream pattern over northern Europe. Kleidon et al. (2000) ran a GCM with a “desert world” and a “green planet” in order to investigate the maximum effect of landscape change. However, these experiments, while useful, do not represent the actual effect of realistic anthropogenic land-use change. Actual documented land-use changes are reported, for example, in Baron et al. (1998), Giambelluca et
al. (1999), Leemans (1999), and O’Brien (1997, 2000). Giambelluca et al., for example, reports albedo increases in the dry season of from 0.01 to 0.04 due to deforestation over northern Thailand.

Figure 8 illustrates how precipitation patterns in the tropics are altered in southeastern Asia and adjacent region in a general circulation model (GCM) where two 10-year simulations were performed: one with the current global seasonally-varying leaf area index (LAI) and one with the potential seasonally-varying leaf area index, as estimated by Nemani et al. (1996). No other landscape attributes were changed. The figure presents the 10-year average difference in precipitation for the month of July for the two GCM sensitivity experiments, which illustrates major pattern shifts in precipitation. As with El Niño, this alteration in tropical thunderstorm patterning teleconnects to higher latitudes as shown in Figure 8b, where the 10-year averaged 500 mb heights for July are presented. The 10-year averaged 500 mb heights are also shown for January (Figure 8a).

The GCM produced a major, persistent change in the trough-ridge pattern of the polar jet stream, most pronounced in the winter hemisphere, which is a direct result of the tropical land-use change. Unlike an El Niño, however, where cool ocean temperatures return so that the El Niño effect can be clearly seen in the synoptic weather data, the landscape change is permanent. Figure 9 shows how the 10-year averaged surface-air temperatures changed globally in this model experiment (Chase et al. 1996).

That landscape change in the tropics affects cumulus convection and long-term precipitation averages should not be a surprising result, based on the discussions earlier in this paper. For example, as reported in Pielke et al. (1999b), using identical observed meteorology for lateral
boundary conditions, the Regional Atmospheric Modeling System was integrated for July-August 1973 for south Florida.

Three experiments were performed - one using the observed 1973 landscape, another the 1993 landscape, and the third the 1900 landscape, when the region was close to its natural state. Over the two-month period, there was a 9% decrease in rainfall averaged over south Florida with the 1973 landscape and an 11% decrease with the 1993 landscape, as compared with the model results when the 1900 landscape is used.

The limited available observations of trends in summer rainfall over this region are consistent with these trends. Chase et al. (2000) completed more general landscape change experiments using the CCM3 from the National Center for Atmospheric Research (NCAR). In this experiment, two 10-year simulations were performed using current landscape estimates and the potential natural landscape estimate under current climate. In addition to LAI differences, albedo, fractional vegetation coverage, and aerodynamic roughness differences were included. While the amplitude of the effect of land-use change on the atmospheric response was less than when the CCM2 GCM model was used, substantial alterations of the trough-ridge polar jet stream still resulted.

Figures 10-12 show the January 10-year averaged cumulus convective precipitation, 200 mb height, and near-surface temperature differences between the two experiments. Despite the difference between the experiments with CCM2 and CCM3, both experiments produce a wave number three change pattern in the polar jet stream. Pitman and Zhao (2000) and Zhao et al. (2001) have recently performed similar GCM experiments and have provided confirmation of the Chase et al. (1996, 2000) results.
Other studies support the result that there is a significant effect on the large-scale climate due to land-surface processes (e.g., Idso et al. 1975; Rodríguez-Iturbe et al. 1991a,b; Entekhabi et al. 1992; Sud et al. 1993, 1995; Xue and Shukla 1993, 1996; Dirmeyer 1994; Foley et al. 1994, 1998; Milly and Dunne 1994; Brubaker and Entekhabi 1995; Claussen 1995, 1998; Entekhabi and Brubaker 1995; Eltahir 1996; Xue 1996, 1997; Xue et al. 1996; Betts et al. 1997; Broström et al. 1998; Ganopólsik et al. 1998; Ferranti and Molteni 1999; Fraedrich et al. 1999; Pitman et al. 1999; Zeng and Neelin 1999; Zeng et al. 1999; Burke et al. 2000; Costa and Foley 2000; Hoffmann and Jackson 2000; Kleidon et al. 2000; Koster et al. 2000; Porporato et al. 2000; Ramirez and Senareth 2000; Texier et al. 2000). Zeng et al. (1998) found, for example, that the root distribution influences the latent heat flux over tropical land. Kleidon and Heimann (2000) determined that deep-rooted vegetation must be adequately represented in order to realistically represent the tropical climate system.

Dirmeyer and Zeng (1999) concluded that evaporation from the soil surface accounts for a majority of water vapor fluxes from the surface for all but the most heavily forested areas, where transpiration dominates. Recycled water vapor from evaporation and transpiration is also a major component of the continental precipitation. Brubaker et al. (1993) found that locally contributed water vapor to precipitation generally lies between 10 to 30%, but can be as high as 40%. Eltahir and Bras (1994) concluded that there is 25-35% recycling of precipitation water in the Amazon. Dirmeyer (1999) concluded that interannual variations of soil wetness are large enough to influence climate GCM simulations.

An important conclusion from such studies is that land-use change directly alters local and regional weather and climate in two ways. First, the local and regional atmospheric conditions are changed since the Bowen ratio is changed as the surface heat, moisture, and other
trace gas and aerosol budgets are altered. Secondly, larger-scale heat and moisture convergence, and associated large-scale wind circulations can be changed as a result of changes in the large-scale atmospheric pressure field due to the landscape change.

These regional and global model studies indicate that the spatial patterning of deep cumulus convection particularly in the tropics and midlatitude summers are significantly altered as a result of landscape changes. These alterations in cumulus convection teleconnect to middle and higher latitudes, which alters the weather. This effect appears to be most clearly defined in the Winter Hemisphere.

6. Conclusions

This chapter demonstrates that vegetation and soil processes and change directly affect the atmosphere and climate globally on a variety of time and space scales. This alteration in fluxes directly modifies the environment for thunderstorms which are an effective conduit for heat, moisture, and momentum to higher latitudes, landscape processes exert a major influence on global weather and climate. In the context of climate, soil, and vegetation dynamics are as much a part of the climate system as atmospheric variables (Hayden 1998; Pielke 1998; Wang and Eltahir 2000a,b). New observational platforms, such as the Tropical Rainfall Measuring Mission (TRMM; Tao et al. 2001), offer opportunities to develop improved understanding of the role of surface-atmosphere interactions on cumulus convective rainfall.

Since atmospheric and ocean circulation patterns and their subsequent involvement within the planet’s climate are dynamic, variable, and difficult to predict, this limits the ability to predict the impact of land-use change and landscape dynamics on climate patterns. As a result, as discussed in Pielke et al. (2002), manipulating land-surface conditions for the purpose of carbon
sequestration under the Kyoto Protocol could have a variety of unanticipated impacts on global and regional climate. The Kyoto Protocol uses only the global warming potentials (GWPs) of the regulated greenhouse gas molecules listed in its Annex A as its mitigation currency. A more complete indication of human contributions to climate change will require the climatic influences of land-surface conditions and other processes to be factored into climate change mitigation strategies. Many of these processes will have strong regional effects that are not represented in a globally-averaged metric. The currency of global and regional human-caused changes in terms of a Regional Climate Change Potential could offer a new metric useful for developing a more inclusive protocol. This concept would also implicitly provide a way to monitor potential local-scale environmental changes that could influence biodiversity.

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References


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Figure 1: Examples of global land-use change from 1700, 1900, 1970, and 1990. The human-disturbed landscape includes intensive cropland (red), and marginal cropland used for grazing (pink). Other landscape includes, for example, tropical evergreen and deciduous forest (dark green), savanna (light green), grassland and steppe (yellow), open shrubland (maroon), temperate deciduous forest (blue), temperate needleleaf evergreen forest (light yellow), and hot desert (orange). Of particular importance in this paper is the expansion of the cropland and grazed land between 1700 and 1900 (from Klein Goldewijk 2001).

Figure 2: Changes in global land cover as a result of human activities, expressed as differences in afternoon radiometric temperatures. These changes represent deviations from ambient surface temperatures under natural vegetation during midsummer, clear sky conditions. Increases result from deforestation while decreases are due to irrigated agriculture (from Nemani et al. 1996).

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will also influence the heat budget (adapted from Pielke and Avissar 1990 with kind permission from Kluwer Academic Publishers).

Figure 4: Schematic illustration of the surface moisture budget over (top) bare soil, and (bottom) vegetated land. The roughness of the surface (and for the vegetation, its displacement height), will influence the magnitude of the moisture flux. Dew and frost formation and removal will also influence the moisture budget (adapted from Pielke and Avissar 1990 with kind permission from Kluwer Academic Publishers).

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Figure 12: Difference in near-surface air temperature (current minus natural landscape) using a 9-point spatial filter for easier visibility. Contour intervals are 0.5, 1.0, 1.5, and 3.0°C. Lighter shading represents the 90% significance level for a 1-sided t-test. Darker shading represents the 95% significance level (from Chase et al. 2000 with permission from Springer-Verlag).
Figure 1: 1900
Figure 1: 1970
Figure 1: 1990
Global Land Cover Changes Expressed as Changes in Afternoon Radiometric Temperature

Figure 2
Figure 3
Figure 4
Figure 5
Figure 6
Figure 10
Figure 11
NEAR SURFACE TEMPERATURE DIFFERENCE

Figure 12
Table 1: Tropical forest extent and loss (rain forest and moist deciduous forest ecosystems)  
(Source: World Resources Institute, 1994; adapted from O’Brien 2000).

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