Vegetation Stress as a Feedback Mechanism in Midlatitude Drought

PAUL A. DIRMeyer

Center for Ocean–Land–Atmosphere Studies, Calverton, Maryland

(Manuscript received 18 May 1993, in final form 11 February 1994)

ABSTRACT

An atmospheric general circulation model with land surface properties represented by the Simplified Simple Biosphere Model is used to investigate the effect of soil moisture and vegetation stress on drought in the midlatitudes. An idealized land–sea distribution with simple topography is used to remove as many external sources of climate variation as possible. The land consists of a single, flat, rectangular continent covered with prairie vegetation and centered on 44°N of an aqua planet. A control integration of 4 years is performed, and several sets of seasonal anomaly integrations are made to test the sensitivity of seasonal climate to low initial (1 April) soil moisture and dormant vegetation like what would occur during a severe drought.

It is found that the inclusion of dormant vegetation during the spring and early summer greatly reduces evapotranspiration by eliminating transpiration. This affects local climate more strongly as summer progresses. Low initial soil moisture, combined with dormant vegetation, leads to a severe drought. The reduction in precipitation is much greater in magnitude than that due to low soil moisture alone, and greater than the sum of the effects computed separately. Although the short-term drought is more severe, the dormancy of the vegetation prevents further depletion of moisture in the root zone of the soil, so soil moisture begins to rebound toward the middle of summer.

1. Introduction

Drought in agricultural regions of the midlatitudes is a long-standing problem for economic interests and weather forecasters alike. The failure of rainfall during the spring and summer months can seriously affect the harvest of many crops, thereby limiting food supplies for man and livestock. Severe droughts, like those of the North American "dust bowl" in the 1930s, can change both the economy and landscape for many years. Although serious widespread droughts are rather rare, droughts of smaller scale affect agricultural areas every year.

Studies of climate have shed some light on the factors that cause and prolong drought. It has been understood for some time that remote effects such as sea surface temperature (SST) anomalies can alter the general circulation over continents in the midlatitudes (Bjerknes 1969). Such changes can increase the likelihood of drought by promoting anomalous subsidence over regions (Trenberth et al. 1988; Lau and Peng 1990). However, numerous modeling studies suggest that such teleconnections alone cannot account for all aspects of observed droughts, such as the North American drought of 1988 (Mo et al. 1991; Trenberth and Branstator 1992).

It has long been speculated that anomalously low soil moisture on local or regional scales may contribute to the initiation or maintenance of drought in the midlatitudes (Namias 1959, 1962). Large aberrations in soil moisture can cause changes in circulation and the distribution of rainfall (Shukla and Mintz 1982). Observational studies have suggested that there is correlation between springtime conditions and summer drought (Namias 1960), and some areas are predisposed to prolonged anomalies in soil moisture and precipitation (Karl 1983). Modeling studies confirm the correlation between antecedent soil moisture and rainfall (Carson and Sangster 1981; Rind 1982; Mintz 1984; Oglesby and Erickson 1989; Oglesby 1991). Dirmeyer and Shukla (1993) present a thorough review of modeling and observational studies linking soil moisture anomalies to changes in climate. Garratt (1993) reviews the various land and boundary-layer models in use, and the sensitivity of climate simulations to these models.

Direct evaporation from the soil is clearly affected by low soil moisture. However, it is not clear how great a role the state of vegetation plays in changing surface fluxes. Typically, when a springtime drought develops, vegetation growth is curtailed by the lack of available soil moisture. This is particularly true of annual crops, grasses, and deciduous plants. If the drought persists into the heat of the summer, the normal growth of leaf area slows or stops (Barlow 1986). Experimental work shows that arable crops or grassland vegetation in the

© 1994 American Meteorological Society
central Great Plains of the United States may stop transpiring five to ten days after the last rainfall (Sellers and Hall 1992). A percentage of the stressed vegetation withers or dies, greatly affecting fluxes of moisture and energy between plants and the environment (Turner 1986; Schulze 1986). To maintain the balance, the magnitude of other terms of the moisture and energy budgets must change. If the changes are large enough, there can be a systematic response in the local climate.

This response of vegetation to drought is not adequately simulated in most biosphere models. Studies of drought have included biological surface models (M. Fennnessy 1993, personal communication), but the annual cycle of vegetation in these models is specified and does not respond to drought in the manner described above. This lack of interannual variability causes modeled vegetation in drought regions to be specified as unrealistically robust, and transpiration, although reduced because of moisture stress, is still probably too large. Parameters such as displacement height, net albedo, and bulk boundary-layer resistance are incorrectly specified as well.

This experiment attempts to determine if the changes in vegetation that occur during a drought can have an effect on the climatology of the drought itself. To elucidate not only the local changes to the fluxes of moisture and energy at the surface, but also the changes in circulation and moisture transport over a large area, a general circulation model (GCM) is used. A control integration is compared to three drought cases with varying initial (1 Apr) soil moisture and vegetation status. Our approach is to use a GCM with simplified boundary conditions and a large, widespread perturbation to the land surface. Our intent is to see if a significant response to changes in vegetation exists on the seasonal timescale, when simulations are performed at GCM resolution. Positive results may justify examining the problem with more realistic scenarios, with the aim of incorporating data on the interannual variability of vegetation into forecast models. No attempt is made in this study to examine how other factors (e.g., soil porosity, albedo, etc.) affect response to drought.

Section 2 describes the GCM and biosphere model used, as well as the design of the experiment. Section 3 summarizes the control integration, to familiarize the reader with the climate of this idealized planet. Section 4 presents results from the anomaly integrations. Discussion is given in section 5, and concluding remarks are provided in section 6.

2. Description of the model

In all experiments, an atmospheric GCM is used with specified lower boundary conditions over ocean and a biosphere model over land to investigate the effects on climate of changes in vegetation and soil moisture.

a. The atmosphere model

The GCM used is a research version of the National Meteorological Center (NMC) Global Spectral Model described by Sela (1980) with modifications, boundary conditions, and initial conditions as described by Kinter et al. (1988). The horizontal truncation is at total wavenumber 15, which is equivalent to a resolution of approximately 4.5° latitude by 7.5° longitude. A Gaussian grid is used by the model in physics calculations. The model is discretized into 18 vertical layers with resolution concentrated near the lower boundary. The time step for integration is 30 minutes.

The radiative processes in the GCM include the diurnally varying radiation scheme of Harshvardhan et al. (1987), the shortwave transfer scheme described by Lacis and Hansen (1974) as modified by Davies (1982), and the longwave scheme of Harshvardhan and Corsetti (1984). Boundary layer fluxes are those used in the E2-Physics package of the Geophysical Fluid Dynamic Laboratory (GFDL) model (Miyakoda and Sirutis 1986). Vertical diffusion in the boundary layer is based on the second-order closure model of Mellor and Yamada (1982). Convection and large-scale precipitation are computed using a modification of the scheme of Kuo (1965) as described by Sela (1980), and the shallow convection scheme is that of Tiedje (1984). The current version of the GCM has interactive model-generated cloud scheme of Hou (1990). Schneider and Kinter (1994) describe the climatology of this version of the model.

b. Simple biosphere model

The biosphere model used to supply the lower boundary conditions to the GCM is a simplified version of the Simple Biosphere (SiB) Model of Sellers et al. (1986), which is described by Xue et al. (1991) and referred to hereafter as SSiB. The original SiB model of Sellers et al. (1986) includes three soil layers and two layers of vegetation: a canopy layer and a ground cover. SSiB retains the three soil layers for moisture calculations (two layers for soil temperature), but combines the two layers of vegetation into one without seriously compromising the simulated fluxes of radiation, heat, moisture, and momentum. The efficiency of calculations of the diurnal variation of albedo and the dependence of stomatal resistance on soil moisture have been greatly improved in SSiB. Also, the mechanical fluxes between canopy and atmosphere have been parameterized.

Each vegetation type is based on a physically observed vegetation variety and has its own set of values for the physical, physiological, and morphological SSiB model parameters. Several of these parameters are specified as a function of time, varying from month to month to simulate the annual cycle of the vegetation, while other parameters are invariant. Detailed expla-
nation of SiB can be found in Sellers et al. (1986), Sellers and Dorman (1987), and Dorman and Sellers (1989). Sato et al. (1989) describes the implementation of SiB as the land surface model coupled to the GCM described earlier.

c. Boundary conditions

In all of the experiments, the atmospheric GCM overlies an aqua planet with a single continent. The continent is rectangular with uniform vegetation type. There are no mountains, nor are there zonal asymmetries in SSTs. There is consequently little energy in stationary and quasi-stationary eddies, which might predispose an area toward drought (Broccoli and Manabe 1992). Also, this approach removes many of the potential nonlinear interactions that could obscure local response to changes at the land surface (Blackmon et al. 1987; Nigam et al. 1988). In such a manner, the regional feedbacks of drought development and maintenance can be isolated as much as possible. Other studies have shown that idealizations can help strengthen the signal of land–atmosphere interactions in GCM experiments (e.g., Manabe and Stouffer 1980; Manabe et al. 1981).

The vegetation type used in all experiments is prairie with properties specified by Dorman and Sellers (1989). This vegetation type is used because of its prevalence in drought-prone areas and its similarity to grain crops as specified in SSIB. No other vegetation types were tested. In two of the cases the vegetation parameters are held at January values throughout the integration. Table 1 gives a list of the values of SSIB parameters used in this study. There are several parameters of particular importance to the experiment. The grass is 0.6 m high, and the roughness length is 0.08 m. There exists a large variation in leaf area index with time. Displacement height varies between 0.22 and 0.33 m with season, and 90% of the surface is covered by vegetation. Vegetation color and leaf area peak in July. Roots penetrate about 0.5 m into the soil. Net albedo varies with the seasons and is dependent on the leaf area index, the ratio of live versus dead plant material, and the zenith angle of the sun. Snow cover also affects net albedo.

The SSTs are prescribed and vary with the seasons. The SSTs are computed from zonal averages of the observed climatological data of the Comprehensive Ocean–Atmosphere Data Set (COADS: Slutz et al. 1985). The data are then averaged together across hemispheres to remove asymmetries for like seasons. In other words, the data for month $m$ at latitude $\phi$ are averaged with the data for month $m + 6$ (or $m - 6$) at latitude $-\phi$, so that the SSTs in the Northern Hemisphere for any season are symmetric about the equator with the SSTs for the same season in the Southern Hemisphere. Thus, the annual cycles in opposite hemispheres are offset by 6 months. Initial soil moisture and temperature are generated by the same process as the SSTs, using the soil moisture data of Willmott et al. (1985) and the soil temperature method of Delsol et al. (1971).

The R15 resolution translates to a Gaussian grid of 40 points in latitude by 48 in longitude. The idealized continent used in all experiments spans six grid points in latitude by eight grid points in longitude, or approximately 27° in latitude by 60° in longitude. Figure 1 shows the geometry and latitude of the continent,

![FIG. 1. Shape and position of idealized midlatitude continent superimposed on coastlines of North America.](image)
Table 1. SiB parameters for the grassland vegetation type.

<table>
<thead>
<tr>
<th>Soil properties</th>
<th>January (dormant)</th>
<th>April</th>
<th>July</th>
</tr>
</thead>
<tbody>
<tr>
<td>Albedo</td>
<td>0.1 (visible)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Porosity</td>
<td>0.2 (near-infrared)</td>
<td></td>
<td>0.42</td>
</tr>
<tr>
<td>Depth of layers</td>
<td>0.6 m</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Saturated hydraulic conductivity</td>
<td>$2 \times 10^{-3}$ m s$^{-1}$</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Soil moisture potential at saturation, $\psi$</td>
<td>$-0.086$ m</td>
<td>7.12</td>
<td></td>
</tr>
<tr>
<td>B factor ($B$) relating soil potential, $\psi$, to wetness, $W(\psi = \psi, W^{-B})$</td>
<td>6.0 m</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Canopy height</td>
<td>0.1 m</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Canopy base</td>
<td>0.5 m</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Rooting depth</td>
<td>$1 \times 10^4$ m$^{-1}$</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Fractional area covered by vegetation</td>
<td>0.9</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Leaf area index</td>
<td>0.7</td>
<td>1.0</td>
<td>4.3</td>
</tr>
<tr>
<td>Green leaf area index</td>
<td>0.4</td>
<td>0.7</td>
<td>3.5</td>
</tr>
<tr>
<td>Roughness length (m)</td>
<td>0.0777</td>
<td>0.079</td>
<td>0.0759</td>
</tr>
<tr>
<td>Displacement height (m)</td>
<td>0.2185</td>
<td>0.2390</td>
<td>0.3252</td>
</tr>
<tr>
<td>Bulk boundary layer resistance (s m$^{-1}$)</td>
<td>22.07</td>
<td>18.38</td>
<td>8.30</td>
</tr>
<tr>
<td>Aerodynamic resistance between ground</td>
<td>24.43</td>
<td>24.96</td>
<td>30.06</td>
</tr>
<tr>
<td>and canopy air space (s m$^{-1}$)</td>
<td>0.10</td>
<td>0.07</td>
<td>0.06</td>
</tr>
<tr>
<td>Visible beam albedo (noon at 40°N)</td>
<td>0.11</td>
<td>0.10</td>
<td>0.08</td>
</tr>
<tr>
<td>Visible diffuse albedo</td>
<td>0.30</td>
<td>0.24</td>
<td>0.29</td>
</tr>
<tr>
<td>Near-infrared beam albedo (noon at 40°N)</td>
<td>0.33</td>
<td>0.35</td>
<td>0.43</td>
</tr>
<tr>
<td>Near-infrared diffuse albedo</td>
<td>0.33</td>
<td>0.35</td>
<td>0.43</td>
</tr>
</tbody>
</table>

compared to real earth geography. The continent has a meridional span between 31.1°N and 57.8°N. The size and latitude of the idealized continent are comparable to most of the midlatitude regions of the land surface, which are typically studied as a unit, such as Europe or agricultural North America. The area of the landmass is roughly $1.4 \times 10^7$ km$^2$. The entire continent is at sea level and is arbitrarily centered on 0° longitude. By limiting the meridional extent of the continent, the use of a single vegetation type can be justified as reasonable, and the role of meridional landsea contrasts can be examined. Similarly, zonal landsea contrasts can be explored by limiting the east–west extent of the landmass. However, the spread is large enough that the boundary forcings on the atmosphere by the continent are apparent at the scales of motion present in the model atmosphere.

d. The experiment design

In this experiment, there is one control integration of four years duration, and three sets of three seasonal drought integrations. The control integration is started from zonally averaged observed conditions from 0000 UTC 1 September 1990 and is integrated for 51 months. Each drought integration consists of three separate integrations initialized from the 0000 UTC 1 April conditions for the control integration in 1992, 1993, and 1994 (the start of months 18, 30, and 42 of the control integration). Since the landsea distribution is not like that of earth, and the SSTs do not represent any single year, the values of the years have no historical climatological significance. Each of the drought integrations is carried out for 120 days.

The distinctions between the control case and the three drought cases are as follows. In the first case, vegetation parameters are held at wintertime values; however, initial soil moisture is not altered. This is called the dormant case. In the other two drought cases, initial soil moisture is set to the lowest values allowed by S SiB. Figure 2 shows the mean 1 April soil moisture (percentage of saturation) of the control integration, and the mean of the reduced values used to initialize the other two drought cases. Soil moisture for the top two soil layers of the control case averages 71%, which equals about 450 mm of water in the top 1.5 m of soil. The reduction of soil moisture in the root zone for the second and third anomaly cases is about 40% everywhere. In the second case, the annual
The lack of topographic variation, along with the zonal symmetry of SSTs, helps to minimize the amplitude of stationary zonal eddies. The mean flow is zonally very symmetric, especially in the Southern Hemisphere where the entire surface is covered by ocean. The stationary eddies that are observed arise only from the contrast between the single flat continent and adjacent waters. The contrasts include differences in surface heat, moisture, and momentum exchange. However, transient eddies in the model (not shown) are no less robust in this configuration of the land surface than with realistic topography.

The land surface has a relatively low effective heat capacity. As a result, the amplitude of the annual cycle of surface temperature is greater over land than ocean. In January, the average temperature at 48°N is 14 K colder over land than over ocean, whereas in July the land is 6 K warmer. Also, the land surface temperature has a pronounced diurnal cycle, which the specified sea surface temperature lacks. Thus, the continent is the primary source of variability in sensible and radiational heating on all timescales. Transient eddies are affected, especially since the landmass lies in the baro-

Fig. 2. Mean soil wetness (percentage of saturation) of root layer on 1 April for (a) control case, contour interval 2%, and (b) initially dry case, contour interval 1%.

cycle of vegetation dictated by SSiB is allowed to proceed unabated. As a result, robust growth of the grass will occur throughout the spring and early summer, despite the extremely low initial soil moisture. This second case is called the initially dry case and is analogous to a situation of drought inception with no vegetation response. In the last case, the vegetation parameters are held at January values throughout the integration. As a result, leaf area and roughness length remain small, and transpiration is negligible. This condition simulates the complete failure of grass or crops. This is referred to as the realistic case, even though the vegetation response prescribed is probably more severe than that observed in an actual drought. Table 2 outlines the four cases. In each drought case, changes are made across the entire continent. Again, this may be an unrealistically large area to be affected by drought at one time. However, such an approach facilitates comparisons of sensitivity to the changes at the lower boundary.

3. The control case

The climatology of this idealized world is understandably different than that observed for the earth.

Fig. 3. Latitude–pressure profile of zonally averaged time mean motion fields from the control case: (a) zonal wind, contour interval 5 m s⁻¹; (b) vertical velocity (omega), contour interval 0.2 mb h⁻¹.
clinic zone of the Northern Hemisphere. This variability, along with the differences in mean heating over land and sea, alter the global circulation.

Unlike the ocean, the land surface is not an unlimited source of moisture. When soil moisture is not near saturation, actual evaporation is below potential evaporation. The simulated prairie vegetation may also limit the release of moisture during dry periods, to conserve moisture for its own use. Air over the continent then becomes drier than over adjacent oceans. When this air is advected over open water by the general circulation, evaporation from the ocean surface may be much greater than what would otherwise occur in the absence of the continent. Thus, the distribution of latent heating in the atmosphere is also affected by the presence of the landmass.

a. Characteristics of the mean flow

Latitude–pressure profiles of the zonally averaged time-mean zonal winds and vertical velocity are shown in Fig. 3. The profile of zonal wind is symmetric about the equator, with maximum westerly velocities of 42 m s$^{-1}$ in each hemisphere. The extent of climatological easterlies at the equator is confined below 700 mb, except for a very small area around 200 mb. Widespread easterlies exist at very high latitudes. The profile of vertical motion is also rather symmetric, but shows the major circulation cells to be present. The descending branches of the Hadley circulation are centered at about 19° latitude, and the ascending branches in the mid-latitudes are at 37°. The weak time-averaged zonal eddies cause the regions of lifting and sinking to be narrower and stronger than for earth.

Time-averaged surface pressure is illustrated in Fig. 4. Although there are some dissimilarities between the two hemispheres, there is not much zonal asymmetry due to the continent. The greatest variations from the zonal mean are 3–4 mb, with a climatological trough over the eastern half of the continent. The surface anticyclones over the poles are a consequence of the lack of sea ice. The meridional distributions of time mean moisture fluxes reveal a dual structure to the intertropical convergence zone with a great deal of symmetry across the equator. Rainfall over the continent is less than the zonal average for those latitudes.

b. Seasonal average

Seasonal averages (April–July) are computed from the last three years of the control integration. Adjustment from initial conditions to model climatology occurs during the first 6–9 months, particularly in soil moisture, so these data are not used. Figure 5 shows the seasonal means of precipitation ($P$), evapotranspiration (ET), and $P - ET$. The domain of the figure is limited to the region of the continent. Precipitation is greatest over the southeast quadrant during this time. Rates exceed 4 mm d$^{-1}$ in some areas. Precipitation generally decreases from south to north across the continent, dropping to near 1 mm d$^{-1}$ at the north coast. Seasonal mean ET is even more zonally symmetric. The ET rates range from less than 2 mm d$^{-1}$ at the northern coast to just under 4 mm d$^{-1}$ at the southern coast.

---

**Table 2. Description of the four cases.**

<table>
<thead>
<tr>
<th>Experiment</th>
<th>Initial soil moisture</th>
<th>Vegetation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Control</td>
<td>normal</td>
<td>seasonally varying</td>
</tr>
<tr>
<td>Realistic</td>
<td>low</td>
<td>perpetual January</td>
</tr>
<tr>
<td>Initially dry</td>
<td>low</td>
<td>seasonally varying</td>
</tr>
<tr>
<td>Dormant</td>
<td>normal</td>
<td>perpetual January</td>
</tr>
</tbody>
</table>

---

**Fig. 4.** Time-mean surface pressure from the control case. Contour interval in the zonal eddy plot is 1 mb.
Calculations reveal that the mean rate of increase of precipitable water during this period is less than 10% of the mean precipitation or evaporation rate during the period and is fairly uniform in distribution. Thus, \( P - ET \) is a reasonably proxy for moisture flux convergence (MFC) on the seasonal scale, especially for exhibiting spatial structure. The resulting field of \( P - ET \) shows that most of the convergence at this time of year is in a zonal band over the center of the continent, although precipitation is greatest over the south and southeast.

c. Seasonal evolution

Monthly means for April through July are presented in this section. All values are calculated by averaging together the monthly means of the last three years of model integration.

A broad trough extends across the continent in the zonal direction throughout the season, with mean low-level westerlies to the south and easterlies to the north. A mean surface cyclone persists at approximately 15°E throughout the period, and the minimum pressure in the cyclone drops from 998 mb in April to 991 mb in July. This drop in pressure is representative of the intensification of land surface heating during this time. In June and July there is a pronounced anticyclone aloft over the continent. The minimum July surface pressure is the lowest attained in any month of the year. The cyclone and trough migrate northward over the period, from 40°N in April to 48°N in July. The tropospheric circulation is quite geostrophic, but with considerable low-level convergence into the cyclone in every month except July. Onshore flow is strong in all months over the southern part of the western coast and the northern part of the eastern coast. There is also onshore flow over the southern coast, particularly over the eastern part. Flow is offshore over the western part of the southern coast in May and June, and precipitation drops noticeably compared to April and July.

Figure 6 shows monthly precipitation rates during the 4-month period. April is the month when mean land surface temperatures in this model begin to exceed SSTs at the same latitude. As a result, April marks the beginning of the convective season over land. Maps of the monthly mean precipitation show more features than the seasonal mean. The predominance of rainfall off the southern coast in April and May is associated with the midlatitude storm track in this model. As summer approaches, the band of rainfall moves north. However, the continent holds the convection in its vicinity by establishing a regional thermally driven circulation. There is a northward progression of convective precipitation as summer arrives. In March the line dividing the area where the majority of rainfall is convective (to the south) from the area to where large-scale precipitation dominates is at the south coast. By April, May, and June it has moved to about 40°N, 45°N, and 48°N, respectively. The northernmost extent occurs in July, when convection is predominant to 50°N. Convection begins receding southward in August.

The maximum in precipitation over the east-central portion of the continent coincides with the position of the cyclone until July, when the pattern changes. In July, the rainband in the south reflects the poleward shift of the storm track. July is also the month when the ITCZ shifts to the Northern Hemisphere. The preponderance of rainfall over the eastern part of the continent corresponds to a broad area of low-level convergence. The northern maximum over land is a manifestation of a secondary zonal precipitation belt, which forms in July and August. Model diagnostics indicate
this precipitation is of large-scale, not convective, origin.

The seasonal progression of ET zonally averaged over the continent is shown in Fig. 7a. Prior to April (not shown), evaporation over the ocean exceeds the ET rate over land at the same latitude. In April, the evaporation rates over ocean and land are comparable. As time progresses, ET over land begins to dominate, and evaporation off the eastern and western coasts is suppressed. By August (not shown), ET over land wanes, and the rates are again comparable at the same latitude over land and ocean.

Precipitation exceeds ET across the central latitudes of the continent during April–June (Fig. 7b). The center of this terrestrial sink of atmospheric moisture is actually off the eastern coast, but a secondary maximum exists just north of the surface cyclone. The sink is flanked by sources of atmospheric moisture to the north and south. These sources strengthen with time until July. There is a conspicuous shift in the pattern in July, which corresponds to the northward shift of the rainband over the southern segment of the continent. The entire central region of the land mass has become a weak source of moisture to the atmosphere. Only at the northern and southern coasts does precipitation exceed ET.

The synoptic situation that contributes to these patterns is as follows. Transient cyclones sweep from west to east across the continent. Although there is a great deal of variation in the tracks and intensities of these systems, most have a weak northward component of motion across the landmass and strengthen as they traverse. There is a strong diurnal component to precipitation associated with these storms, with maximum rainfall typically occurring during early afternoon. The region of low pressure over the eastern half of the continent represents a preferred region for these systems to stall. The average period of these systems is three to five days.

The role the diurnal cycle of ET in supplying precipitable water is of particular importance. During the morning and early afternoon, ET is vigorously supplying moisture to the lower troposphere. The increased flux into the local atmosphere is so great that it can support increased precipitation with water vapor left over for export. The greatest rate of moistening of the lowest 200 mb occurs around 1000 LST. By afternoon, the combination of drying soil, waning shortwave radiation, and increased plant stomatal resistance curtails ET. Moisture that is not advected away is carried aloft by convective processes, so that the peak moistening rate in the midtroposphere occurs around 1400 LST.

Fig. 6. Mean monthly precipitation rates for the control case. Units are mm d$^{-1}$. 
This moisture helps supply precipitation many hours after ET has begun to decline.

4. The drought cases

In this section, the anomalies in the seasonal (Apr-Jul) means, monthly means, and diurnal cycles of the three test cases are examined. The period is meant to represent a typical growing season in the midlatitudes.

a. Seasonal average

The differences between the drought and control cases for precipitation are given in Fig. 8. The last panel shows the difference between the realistic case and the control case. Drought conditions are seen to exist over nearly the entire continent. Rainfall deficits generally extend well beyond the coastlines. The mean negative rainfall anomalies of the initially dry case (second panel) are less than half that of the realistic case and lack the well-defined structure and extent. Nevertheless, a large fraction of the continent shows a precipitation shortfall of 0.6 mm d\(^{-1}\) or more, which is equivalent to a deficit of 72 mm or approximately 20% over the course of the growing season. The first panel shows the anomalies for the case with dormant grass. Anomalies here are weak; 10% or less of the rainfall in the control case. Stronger anomalies of both signs exist offshore, suggesting that the differences in seasonal rainfall over land may not be of significance. It is worth noting that the sum of the anomalies of the initially dry case and the dormant case falls well short of the anomaly when the two effects are combined. The implication of this nonlinear response is that a positive feedback mechanism may work to amplify the drought in the realistic case.

As a check of the significance of these anomalies, compared to the natural variability observed in the

FIG. 7. Mean monthly rates of (a) ET and (b) P – ET, zonally averaged over the continent. Units are millimeters per day.

FIG. 8. Seasonal mean anomalies in precipitation rate: (a) dormant grass case, (b) initially dry case, and (c) realistic case. Contour interval is 0.2 mm d\(^{-1}\).
control integration, the ratio of the mean square difference of monthly mean precipitation anomalies for drought and control runs has been compared. This is formulated as a ratio

$$\frac{\sum (P_a - P_{clim})^2}{\sum (P_{ctl} - P_{clim})^2}$$

where $P_{clim}$ is the control climatological monthly precipitation, $P_{ctl}$ is the monthly control precipitation, and $P_a$ is the monthly precipitation for the given anomaly case. Summations are taken over April–July of the last three years of the integrations, and the ratio is computed over each continental grid point. We find that in the dormant grass case, April–July (AMJJ) anomalies in precipitation are significantly larger (at the 95% confidence level, by Fisher’s $F$ test) than anomalies due to natural variability across 46% of the continent. For the initially dry case the area of significance is 73%, and for the realistic case it is 94%. Thus, the anomalies are significantly greater than what would be expected by chance.

Figure 9 shows the anomalies in ET. In the realistic case, the signature of the continent is again very clear.

The patterns of rainfall and ET anomalies in the realistic case appear to have rather different shapes. However, an interesting similarity appears if one plots the anomalies as a percentage change (Fig. 10). The patterns and magnitudes of the percentage anomalies over land are very similar. In both precipitation and ET, deficits exceed 60% in an area near the northwestern corner. The deficits taper off gradually to the east and somewhat more sharply to the south and north. Evaporation in this case is very strongly limited by a lack of precipitation. This results in a strong positive correlation between the two fields. One does not see such strong pattern correlations in the other two cases (not shown). In fact, the correlation drops as the magnitude of the anomaly drops. If one examines the anomalies over ocean to the north of the continent, one can see a negative correlation between precipitation and evaporation. This is true of the oceanic anomalies in every case because over the ocean, moisture availability is not a limiting factor for evaporation.

Whereas the shape of the anomalies in precipitation and ET tended to vary from case to case, the change in $P - ET$ maintains the same pattern in each case. Figure 11 shows the seasonal anomalies in $P - ET$. The anomalies in $P - ET$ are positive at the northern and southern coasts, and negative between. This pattern over land is out of phase with the field of $P - ET$ in the control case. Thus, it reflects a reduction in moisture flux convergence along the central band, with less divergence to the north and south. In fact, the magnitude of the anomaly in the realistic case is large enough to largely eliminate variations in MFC over the continent.

Table 3 shows the seasonal budget of moisture at the surface, averaged over the entire continent for each case. Also shown for comparison are the values computed by Rasmusson (1968) for eastern North Amer-
ica, averaged over the same four months. Precipitation is greatest in the control case, followed by the dormant grass case, the initially low soil moisture case, and the realistic case. Evapotranspiration decreases in the same order. The continent is actually a source of atmospheric moisture during this period in the control case. In the realistic case, the continent is a distinct sink of atmospheric moisture. The balance is nearly perfect in the other two cases. Runoff is seen to be very low in the two cases with low initial soil moisture, and fairly high otherwise. As a result, the change in soil moisture is seen to be rather small in the two dry cases. This season is normally a time of drying of the soil, as can be seen in the control case. The two cases with initially low soil moisture simply have little moisture left for evapotranspiration. The soil in the realistic case appears to get a little wetter with time. The soil of the dormant case dries like that of the control case, but not as much since transpiration is not removing additional water from the root zone. The values from Rasmusson (1968) show less ET and more moisture convergence, possibly due to the differences in geometry and vegetation between eastern North America and the idealized continent.

**Table 3.** Time-averaged (AMJJ) surface moisture variables. All units in millimeters per day.

<table>
<thead>
<tr>
<th></th>
<th>Control</th>
<th>Dormant grass</th>
<th>Low ISM</th>
<th>Realistic</th>
<th>Rasmusson (1968)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Precipitation</td>
<td>2.97</td>
<td>2.78</td>
<td>2.54</td>
<td>2.02</td>
<td>2.57</td>
</tr>
<tr>
<td>Evapotranspiration</td>
<td>-0.07</td>
<td>-0.01</td>
<td>+0.02</td>
<td>+0.22</td>
<td>+0.30</td>
</tr>
<tr>
<td>$P - E$</td>
<td>0.73</td>
<td>0.66</td>
<td>0.17</td>
<td>0.15</td>
<td>0.65</td>
</tr>
<tr>
<td>Runoff</td>
<td>-0.80</td>
<td>-0.67</td>
<td>-0.15</td>
<td>+0.07</td>
<td>-0.35</td>
</tr>
<tr>
<td>Soil moisture change</td>
<td>-0.80</td>
<td>-0.67</td>
<td>-0.15</td>
<td>+0.07</td>
<td>-0.35</td>
</tr>
</tbody>
</table>
Precipitation recycling has been computed following the method of Brubaker et al. (1993). The recycling ratio (percentage of rainfall over area A that originated as evaporation from area A) is computed using their Eq. (16) for the entire continent. The seasonal mean ratio is 0.76 for the control case, 0.73 for the dormant grass case, 0.72 for the initially dry case, and 0.64 for the realistic case. Thus, the realistic case reflects how the lack of available moisture over land impacts precipitation. The ratio peaks in every case during May or June, although the maximum difference between any two months within a case during the season is 0.09.

Table 4 lists a budget of the seasonal mean values of a number of surface variables averaged over the continent. Mean cloud cover is not greatly affected by the changes in surface properties. Even in the realistic case, where precipitation is reduced dramatically, mean cloud cover remains about the same. However, this does not mean changes do not occur in the distribution of cloudiness. Downward shortwave radiation does increase, particularly in the realistic case.

The net surface albedo is slightly lower for dormant vegetation. This occurs as a result of the differences in optical parameters between dead and healthy grass as specified in the SSIB model, and the amount of soil exposed due to changes in the leaf area index. It appears that dormancy diminishes albedo by roughly 0.014. As a result, less shortwave radiation is reflected in the drought cases. The combination of increased downward shortwave radiation and decreased albedo means more shortwave radiation is absorbed. The warming of the surface soil in each of the drought cases exhibits this effect.

Net energy absorbed into the ground, computed as a residual of the surface energy balance, shows that the soil of the control case actually retains the most energy of all cases. Yet the surface and deep soil temperatures of the control case are the lowest. Likewise, the dormant grass case ranks second in energy absorption and coolness of soil. The reason for this apparent discrepancy can be found by examining the soil moisture. In the control and dormant cases, soil moisture is high. The energy absorbed by the soil is spent heating wet soil, which has a much greater heat capacity than dry soil. The discrepancy in soil moisture is also reflected in the difference between the surface and deep soil temperatures. The difference in the two cases with high soil moisture is ≈ 0.4 K, while for the dry cases the difference is ≈ 1.0 K. Wet soil also has a higher thermal conductivity, which allows heat at the surface to penetrate to depth more readily.

Aside from the obvious differences in soil wetness based on the initial soil moisture, there are distinctions between dormant and healthy vegetation cases. Soil moisture for the dormant grass case is greater than the control case. Likewise, soil in the realistic case is wetter than the initially dry case. This difference is largest in the root zone. In the cases where the vegetation is healthy, transpiration is removing moisture from the root zone. The vertical gradients formed by the deficit in this layer draw moisture from adjacent layers, drying them slightly as well.

The availability of soil moisture clearly affects latent heat flux from the surface. Latent heat flux is highest when soil moisture is abundant, and both transpiration and evaporation are available mechanisms for transporting water from soil to atmosphere. The percentage of total surface heat flux accounted for by ET, which is sometimes called “evaporative fraction” (Shuttleworth et al. 1989; Sellers and Hall 1992) and computed as LH/(SH + LH), where LH is latent heat flux and SH is sensible heat flux, is 64% in the control case. When the transpiration mechanism is removed in the dormant case, the ratio drops to 58%. In the initially dry case (with transpiration), it is 53%, and in the realistic case, it is only 38%. The sum of latent and sensible heat flux is quite uniform in each case: 137–140 W m⁻². So latent heat flux can be said to occur when it can. When moisture is not available, sensible heating approximately makes up the difference.

Downward longwave radiation at the ground is also approximately constant between cases. However, longwave radiation emitted back to the atmosphere varies with surface temperature. As a result, net longwave radiation for each case ranks in the same order as evaporation fraction. The dry cases radiate more thermal energy, since dry soil cannot effectively retain or conduct that energy.

### Monthly progression

Figure 12 shows the seasonal evolution of eight different variables for each case, averaged over the entire
continent. Values are ensemble means among the three integrations in each case. The variables shown are precipitation, ET, runoff, wetness of the soil’s root zone, surface temperature, sensible heat, evaporative fraction, and MFC. Runoff is defined as the precipitation not intercepted by vegetation nor absorbed into the soil. Runoff is removed from the surface after each time step of the integration, so there is no puddling or tracing of water into streams and rivers.

In April, the vegetation in all cases is almost inactive, so the major grouping in all surface variables is between the two initially dry cases (the initially dry case and the realistic case) and the two cases with unaltered soil moisture (the dormant case and the control case). For the initially dry cases, precipitation, ET, runoff, soil moisture, and evaporative fraction are low; surface temperature and sensible heat are high. As time progresses, discrepancies between the state of vegetation in the paired cases begin to manifest themselves. By May, the two initially dry cases have begun to diverge. In the realistic case, evaporation and precipitation do not increase with time as in the other cases. However, sensible heat and temperature do increase. The two cases with nominal soil moisture are still very similar in May, although discrepancies in surface temperature are beginning to appear.

By June, the dormant case has noticeably digressed from the control case. Precipitation, ET, and evaporative fraction are increasing more slowly, runoff and soil wetness have begun to decrease, temperature increases more rapidly, and sensible heating fails to decrease. At the same time, the two initially dry cases have continued to diverge, so that by June, the dormant case and the initially dry case are comparable, while
the case with both dormant vegetation and initially low soil moisture shows a large deviation from the control. This pattern continues into July.

MFC shows less distinct stratification between cases than the surface variables. However, MFC for the realistic case has much less variation from month to month and is generally of a smaller magnitude. This is consistent with the weakening of the patterns of $P - E$ shown in Fig. 11.

The progressive effect of dormant vegetation on local climate is evident in the precipitation anomalies of the dormant case. Figure 13 shows the mean monthly precipitation anomalies (dormant minus control) for each month. In April and May, anomalies of both signs are evident over the continent with approximately equal extent. The magnitude of the anomalies generally do not exceed 10% of the control values. By June, negative anomalies begin to dominate in extent, and the deficit over much of the eastern continent exceeds 20%. By July, virtually the entire continent is covered by negative anomalies, with shortfalls in the northeast exceeding 1.4 mm d$^{-1}$, or about one-third of control rainfall. Thus, the seasonal average underestimates the final effect of dormant vegetation. A similar progression is found in the realistic case (not shown), where negative anomalies are widespread in each month, but the peak magnitude of the deficits grow from 30%–35% in April to 75%–80% in July. The largest percentage reduction in zonally averaged rainfall shifts from 44°N in April to 51°N by July.

It should be noted that while dormant vegetation seems to intensify the climatic crisis; it also helps alleviate the presumed local cause of the drought—low soil moisture. In Fig. 12d, soil wetness of the root zone is not as depleted when vegetation is thriving. This distinction between the initially dry case and the realistic case is particularly clear by the end of the season. The lack of transpiration by dead grass prevents further desiccation of the soil. Hence, the withering of vegetation may help curtail long-term persistence of the drought even while exacerbating the short-term anomaly. The percentage of the April soil moisture deficit erased by July is calculated for the realistic case and the initially dry case. This quantity is computed as

$$
\left(1 - \frac{D_j - C_j}{D_a - C_a}\right) \times 100,
$$

where $D$ stands for the soil water content of the drought case, $C$ is the same for the control case, and the subscripts $a$ and $j$ stand for mean April and July quantities. A value of zero means the magnitude of the soil moisture anomaly is unchanged from April to July. A value of 100 means there is no anomaly remaining in July, while negative values mean the magnitude of the dry anomaly has increased. Only the top two layers of the soil are included in this calculation. This quantity is

![Fig. 13. Mean monthly precipitation anomaly of the dormant grass case. Contour interval is 0.2 mm d$^{-1}$.](image-url)
illustrated in Fig. 14. Although deficits continue to be strong across the midcontinent in July, the realistic case shows more recovery in almost all areas. In fact, the soil moisture anomaly is eliminated in the south. In the initially dry case, the soil moisture deficit actually worsens in the central and northeastern regions, and only modest gains are made along the southern coast.

The effect of the land surface anomalies on the progression of individual synoptic-scale weather systems is shown in Fig. 15. A succession of storms traveling across the continent is shown in a Hovmöller diagram at 47°N for one particular year of the control and realistic integrations. Mean daily precipitation rates are plotted: darker shading indicates higher rates. Also shown are the tracks of individual storms, defined as minima in the time series of surface pressure. Diurnal variations show up in the plot as a string of dark spots in the track of each system, signifying the location of the system during each afternoon. There is clearly a decrease in both the number of rain events, and in their extent, in the realistic drought case. It seems that the period of the events has not changed. What has changed is the number of synoptic systems that bring substantial rain. Many systems bring no appreciable rainfall. Very few systems are able to spread a band of precipitation across the entire continent in the realistic case, whereas most of the systems in the control case do. Nonetheless, when rain does occur, it appears to be as intense as in the control case. This shows that the reduction in ET does not uniformly reduce the rate of precipitation within synoptic-scale systems, but reduces both the number of systems that produce substantial rain and the lifetime of the rain events.

Figure 16 shows the diurnal cycle of anomalies in precipitation, ET, MFC, and rate of change of precipitable water zonally averaged over the continent for the realistic case minus control during July. Anomalies in the diurnal cycle of the realistic case show that precipitation rates are lower at all hours in the north. There is a clear diurnal variation in the anomaly field in the south. Overnight precipitation is nearly unchanged, but afternoon rainfall, which is primarily convective, is significantly reduced. The magnitude of the anomaly grows as one leaves the southern coast. A similar pattern is seen in ET. The removal of the mechanism for transpiration reduces daytime surface latent heat flux. The latitude of maximum reduction is north of the largest precipitation anomaly and is mirrored in precipitable water. This reemphasizes that local evaporation is an important source of atmospheric moisture over the continent. The lack of ET is only partially offset by increased MFC in the north, mainly supplied by moisture drawn from the south. In the control case, ET increases precipitable water during the morning and early afternoon, and precipitation removes it in the evening and overnight. The magnitude in the diurnal cycle of precipitable water is reduced in the realistic drought case. These anomalies are evident in April and grow in magnitude through July.

Lesser effects are seen in the other two anomaly cases (not shown). In the cases with dormant vegetation, a change in low-level meridional wind suggests that the reduction in surface friction caused by the lack of vegetation allows moist southerly winds to penetrate further into the continent at all hours, shifting the region of convergence northward and depriving the southern regions of rain. Afternoon rainfall still declines, but the paucity is greatest near the coasts where drag-induced convergence has deteriorated. These results are similar to those found by Xue and Shukla (1993) in the simulation of desertification over western Africa.

5. Discussion

We have examined drought situations over land, where precipitation is reduced during the spring and summer. The processes examined entail local changes at the land surface that directly reduce evaporation or transpiration. This results in a decrease in the supply of moisture in the local atmosphere that is available for precipitation. Changes at the surface can also have
a secondary affect by altering the general circulation, thus modifying the distribution of MFC. Although the scenario is idealized, the mechanisms involved are no different than in realistic simulations.

In general, MFC supplies precipitation directly by concentrating the abundant moisture in the lower atmosphere. The correlation between the two is always very high. Both large-scale (dynamically driven) precipitation, and convective precipitation are associated with MFC. In the former, lifting is directly attributable to low-level convergence. Given the spatial and temporal scale of dynamical systems, much of the moisture that precipitates in these systems may travel great distances and be airborne for many days. In the latter case, concentrated diabatic heating drives the lifting, and continuity demands that low-level convergence fill the void. The scale of the convective process is much more limited in space and time, and a greater share of the moisture involved is likely to be of local origin.

The connection between ET and precipitation is more tenuous, but no less important. Evapotranspiration supplies the atmosphere with water vapor, which may then be concentrated by convergence. Thus, the source of terrestrial moisture and the region of precip-
itation may be removed from each other in space and time. As the months progress from spring to summer, the link between ET and precipitation becomes more direct. Local ET accounts for a greater share of the water budget as convection becomes predominant, and large-scale dynamics diminish in strength and importance.

One would expect to see this correlation in the anomalies as well, especially since the strength of the ET anomalies also increases from April to July. Table 5 shows the correlation coefficients for the anomalies in precipitation versus anomalies in ET and MFC over the continent as a function of month. Three-year means at all 48 grid points over land are used to compute the correlations. However, there are far fewer than 48 degrees of freedom (DOFs). It is likely that there are approximately 7–9 DOFs, although the actual number is somewhat variable, depending on the predominant scales of the circulation at any given time. For this table, 9 DOFs are assumed, for which \( r \geq 0.58 \) is significant at the 5% level. These values are shown in bold type. Additional anomalies are compared as well. There are two dormant vegetation anomalies: dormant minus control (effect of dormancy in a well-watered situation), and realistic minus initially dry (effect of dormancy when soil is dry). Likewise, there are two anomalies for low soil moisture: initially dry minus control (effect of dry soil in the presence of active veg-

<table>
<thead>
<tr>
<th></th>
<th>Apr</th>
<th>May</th>
<th>Jun</th>
<th>Jul</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>( P ) vs ET</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Dormant-Control</td>
<td>0.11</td>
<td>0.20</td>
<td>-0.02</td>
<td>0.48</td>
</tr>
<tr>
<td>Realistic-Initially dry</td>
<td>0.58</td>
<td>0.78</td>
<td>0.89</td>
<td>0.75</td>
</tr>
<tr>
<td>Initially dry-Control</td>
<td>0.37</td>
<td>0.25</td>
<td>0.14</td>
<td>0.79</td>
</tr>
<tr>
<td>Realistic-Dormant</td>
<td>0.23</td>
<td>0.52</td>
<td>0.81</td>
<td>0.75</td>
</tr>
<tr>
<td>Realistic-Control</td>
<td>0.09</td>
<td>0.44</td>
<td>0.30</td>
<td>0.59</td>
</tr>
<tr>
<td>( P ) vs MFC</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Dormant-Control</td>
<td>0.74</td>
<td>0.55</td>
<td>0.68</td>
<td>0.44</td>
</tr>
<tr>
<td>Realistic-Initially dry</td>
<td>0.61</td>
<td>0.61</td>
<td>0.70</td>
<td>0.54</td>
</tr>
<tr>
<td>Initially dry-Control</td>
<td>0.74</td>
<td>0.84</td>
<td>0.86</td>
<td>0.83</td>
</tr>
<tr>
<td>Realistic-Dormant</td>
<td>0.70</td>
<td>0.80</td>
<td>0.62</td>
<td>0.50</td>
</tr>
<tr>
<td>Realistic-Control</td>
<td>0.76</td>
<td>0.82</td>
<td>0.32</td>
<td>0.72</td>
</tr>
</tbody>
</table>
The diurnal cycle of moisture during drought is characterized by reduced daytime evapotranspiration rates because of the lack of available soil moisture or removal of the mechanism of transpiration that transports subsurface moisture to the atmosphere. This reduction limits daytime moisture flux divergence over land and the recharge of precipitable water. Precipitation is affected both locally and in neighboring areas that normally supplied by moisture advected from land areas.

It is clear that changes in vegetation in response to springtime drought conditions can have feedback effects on the atmosphere. Temporary loss of vegetation may heighten the short-term climatic effects of drought, while helping to more rapidly eliminate the original soil moisture deficit. By neglecting the effect of drought-stressed vegetation in land-surface models, the severity of droughts may be underestimated and their duration may be overestimated. Although this study examines only response to large (continental) scale drought, the results are applicable on any scale that is large enough that it can force anomalies in the general circulation. No attempt has been made to address how vegetation changes might interact with remote potential causes of drought, such as quasi-stationary eddies forced by SST anomalies.

Changes in circulation that lead to reductions in rainfall at times when soil moisture is plentiful, or when vegetation growth is not great, may not have much of a reciprocating effect on the circulation. In such instances, vegetation feedback would be minimized. However, the accentuation of drought caused by dead or dormant vegetation could conceivably exacerbate the drought enough to extend beyond the seasonal timescale, provided the timing was right, and the area was predisposed to large fluctuations in the climate regime (Entekhabi et al. 1992). It also can be seen from this experiment how widespread degradation in vegetation could alter regional climate. The extent of climate change depends largely on the dependence of rainfall on moisture supplies of local origin, and the vagaries of the general circulation.

Acknowledgments. The author would like to thank those who helped in the support and evaluation of this work, particularly Prof. J. Shukla and Dr. P. J. Sellers for their valuable comments. All figures were prepared using the Grid Analysis and Display System (GrADS), and Mr. B. Doty provided much assistance in the development and display of the graphics. This research was supported by National Science Foundation Grant ATM-90-19296. Computer funding was supplied by National Atmospheric and Space Administration Grants NAGW-2661 and NAGW-1269. Additional computer time was provided by the NASA Center for Computational Studies. Parts of this report have been used in a Ph.D. dissertation at the University of Maryland at College Park.

REFERENCES


Miyakoda, K., and J. Sirutis, 1986: Manual of the E-Physics. [Available from the authors at Geophysical Fluid Dynamics Laboratory, Princeton University, P.O. Box 308, Princeton, NJ 08542.]


