OBSERVATIONAL AND MODELING STUDIES OF THE INFLUENCE OF SOIL MOISTURE ANOMALIES ON ATMOSPHERIC CIRCULATION (REVIEW)

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1. Introduction

The annual mean rainfall for the global continents is estimated to be about 764 mm of which approximately 40% runs off into the oceans. Assuming no secular trends in the annual mean global soil moisture, this suggests that the annual and global mean evaporation from the land surfaces alone is about 60% of the annual and global mean precipitation over the land (Eagleson, 1991). The percentage is even higher during the local summer and in the tropics.

Some of the evaporated moisture may condense and reprecipitate locally. One region where this local water cycle is strong is northern South America. Several independent water balance calculations for the Amazon basin show that approximately 55% of the precipitation there is accounted for by local evapotranspiration (see Salati and Nobre, 1991 for review). On the other side hand, about 75% of the precipitation which falls on the basin is reevaporated before reaching the oceans as runoff (Dickinson, 1991). Figure 1 shows the hydrologic cycle over the Amazon basin.

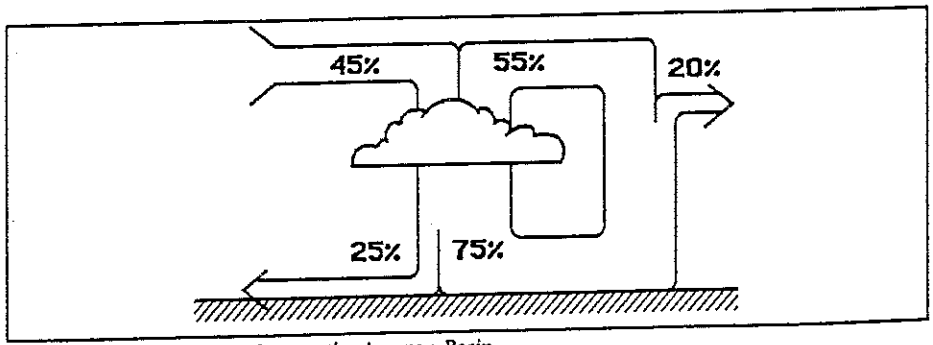

Figure 1. Hydrologic cycle over the Amazon Basin.

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Evaporation from the land surface is a very important component of the global water budget and hydrological cycle. However, it does not necessarily follow that the water evaporated from the land is important in determining the rainfall over land. For example, all the water evaporated from the land could be advected away to the oceans before it recondenses and rains. In that case it will affect the moisture budget and evaporation only over the oceans, which in turn will, of course, affect the moisture supply for rainfall over the land. In order that evaporation from the land affects the rainfall over the land, it is necessary that the prevailing dynamical circulation be such that the land-evaporated moisture recondenses as rain before being advected away. This will depend upon the geographical location of the region under consideration, the prevailing advective velocity, the structure and intensity of the convergence field, and the vertical distribution of moist static energy which determines the nature of the moist convection.

In nature, the total rainfall averaged over the appropriate space and time scales is determined by the combined effects of available and precipitable moisture, and the character and intensity of the dynamical circulation necessary to lift the moisture for condensation and precipitation.

The role of soil moisture is twofold. First, it determines the rate of evaporation, and therefore the moisture supply. Second, it influences the partitioning of incoming radiative energy into sensible and latent heating. Soil wetness influences the heating of the ground which determines the sensible heat flux and affects the dynamical circulation by generation or dissipation of "heat lows". The interaction between the heat lows generated by solar heating of the ground and the associated circulation and rainfall is further complicated by the fact that the maintenance and the intensification of the dynamical low is largely influenced by the latent heat of condensation. For example, if the soil is saturated with water, and the evaporation is equal to the potential evaporation, there will be maximum possible supply of moisture to the atmosphere. Whether this will increase the rainfall or not will depend upon the nature of dynamical circulation and its associated flow patterns. If the rate at which moist static energy is advected away from the region is larger than its accumulation rate, it will not lead to any increase in rainfall. In the reverse case, it will. For the other extreme situation, when the soil is completely dry and there is no evaporation from the land, there may be a reduction in the rainfall. However, if the heating of the land produced intense low pressure

areas which can converge moisture from the surrounding oceans, the rainfall may not necessarily decrease, and if the convergence of moisture is large enough it may even increase the rainfall. This mechanism will cease to operate once the rain starts falling because the soil will no longer be dry.

When the soil does contain water, it acts as a shallow, but widespread, reservoir. Compared with the oceans, it is a highly variable and inconsistent source of moisture for the atmosphere. The rate of evaporation depends not only on the availability of moisture in the uppermost layers of the soil, but on characteristics of the soil itself, the type and distribution of vegetation rooted in the soil, and ambient conditions in the atmosphere near the surface of the earth.

Soil moisture is an important link in two of the principal cycles or feedback loops of the earth climate system — the water cycle and the energy cycle. Perhaps most apparent is the role the land surface holds in the hydrologic cycle. Moisture evaporates from the soil, increasing atmospheric humidity and eventually condensing into clouds where it may precipitate back onto the earth's surface. Less obvious is the role soil moisture plays in the energy cycle. The evaporation of soil moisture constitutes a flux of latent heat into the atmosphere. Thus, the availability of soil moisture is a strong control on temperature and the partitioning of energy at the surface. Changes in the energy balance affect atmospheric temperature, and thus, a host of other components of the climate including evaporation itself.

In this review, we will examine some observational evidence that soil moisture fluctuations do indeed affect climate over seasonal time scales. A few of the many computer modeling studies also will be reviewed. Special attention will be given to those studies which explore the use of soil moisture in the prediction of climate.

2. Observational Studies

In this section we review the observational evidence that soil moisture fluctuations affect atmospheric circulation and rainfall at seasonal time scales.

2.1 Sensitivity to soil moisture

Although the importance of soil moisture to the generation of precipitation was stated at least as early as 1935 by E. P. Stebbing (Anthes, 1984), Namias (1959, 1960) was perhaps the first to directly address the problem of soil wetness as a boundary forcing for the atmosphere. He examined monthly precipitation and temperature data for the Great Plains of the United States. The data were seasonally averaged over periods of 60-84 years, depending on availability of the data. Namias then constructed contingency tables with three categories each (normal, above normal and below normal). Tables relating summer temperature to both temperature and precipitation from the antecedent spring show a clear tendency for dry springs to be followed by hot summers, and wet springs to be followed by cool summers (see Table 1). Also, a tendency for persistence of anomalous temperatures from spring to summer was evident.

Table 1.

Western Plains		Subsequent summer temperature			
Spring Temp.	Precip.	Cold	Normal	Warm	Total
Cold		*101*	*70*	*40*	
	Light	29	21	10	*60*
	Moderate	31	18	19	*67*
	Heavy	**41**	31	11	*83*
Normal		*53*	*74*	*81*	
	Light	12	18	34	*64*
	Moderate	18	33	27	*78*
	Heavy	23	23	19	*65*
Warm		*57*	*65*	*87*	
	Light	9	27	**50**	*86*
	Moderate	18	22	22	*62*
	Heavy	30	16	16	*62*

from Namias (1960)

Walsh et al. (1985) have found that errors in surface temperature forecasts for the United States during 1947-80 correlated well to soil moisture anomalies. The mean errors of temperature vary by 0.5-0.7°C over most of the Great Plains and Rocky Mountains according to the sign of the anomaly of soil moisture. About half of the 61 stations included in the study had differences in composited mean specification which were statistically significant at the 95% confidence level for the summer months when errors were categorized according to the sign of the soil moisture anomaly.

Namias (1960) found that for summer precipitation, just as wet/dry springs tend to precede cool/warm summers, cool/warm springs precede wet/dry summers (see Table 2). Especially strong was the tendency for a warm dry spring to usher in a dry summer. Barnston and Schickedanz (1984) found statistical evidence that on a smaller scale, irrigation may increase precipitation. This seems to be especially true when there was low-level mesoscale convergence over the irrigated area.

Table 2.

Western Plains		Subsequent summer precipitation		
Spring temp.	Precip.	Below normal	Near normal	Above normal
Cold		*53*	*73*	*85*
	Light	12	18	30
	Moderate	19	24	25
	Heavy	22	31	30
Normal		*70*	*73*	*65*
	Light	28	17	20
	Moderate	27	26	26
	Heavy	15	30	19
Warm		*87*	*63*	*58*
	Light	49	22	14
	Moderate	24	16	22
	Heavy	14	25	22

based on Namias (1960)

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...nt summer temperature		
...rmal	Warm	Total
...0	*40*	
...1	10	*60*
...3	19	*67*
...1	11	*83*
...4	*81*	
...8	34	*64*
...3	27	*78*
...3	19	*65*
...	*87*	
...	**50**	*86*
	22	*62*
	16	*62*

from Namias (1960)

Namias concluded that, "...moist soil may serve as a cooling reservoir by using for vaporization some of the heat normally associated with the spring to summer building of the upper level anticyclone...," (Namias, 1959), and "...desiccating warm and dry weather over the Plains in spring provides a healthy environment for the lodgement of the upper level anticyclone in the following summer," (Namias, 1960). More recently, Namias (1989) asserted that low springtime soil moisture was a factor in the US drought of 1988, and a recent study by Fennessy and Shukla (1992) supports this possibility. Any effects that soil moisture would have on the atmosphere should occur at the decay time scale of soil moisture. This has been found by Vinnikov and Yeserkepova (1991) to be typically 2-3 months for a 1 m depth of soil. Modeling studies by Carson and Sangster (1981), Rind (1982) and Delworth and Manabe (1988), among others, have implied similar decay time scales.

2.2 Soil moisture as a predictor

Namias recognized that soil moisture anomalies could aid in the persistence of atmospheric circulation anomalies. In particular he examined two case studies (Namias,1959), one involving heavy spring rains over Texas, and one drought over the eastern United States. In the first case, it was found that from February through mid-June, statistical predictions of 700 mb height for consecutive two-week periods were consistently too high over the region. He associated the anomalous trough over the area to the moist soil acting as a cooling reservoir which impeded the normal building of the summertime upper-level anticyclone in that area. In the instance of the drought over the eastern seaboard, July surface temperatures were abnormally high, even though the region was under an area of anomalously low heights and cold advection at 700 mb.

Can soil moisture be used effectively as a prognostic tool? Karl (1983, 1986) has shown that soil moisture indices can aid in long-range forecasting, particularly in spring and early summer. Karl inspected monthly averaged temperature and precipitation for the United States from 1895-1981. From this he computed the Palmer Drought Severity Index (PDSI) (Karl, 1983), moisture anomaly index Z and water content parameter WC (Karl, 1986). All are derived from the Palmer Drought Model (Palmer, 1965). The moisture anomaly index is where K is a spatially and temporally

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$$Z = (P-\hat{P})K ,$$

dependant weighting factor used to standardize the index. P is actual precipitation, and

$$\hat{P} = E\hat{T} + \hat{R} + R\hat{O} - \hat{L}$$

The circumflex indicates climatically appropriate quantities for the existing conditions. ET is evapotranspiration, R is soil moisture recharge, RO is runoff, and L is monthly soil moisture loss. The water content parameter is computed as:

$$WC = W_s + W_u .$$

where s denotes the surface soil layer, and u is the underlying layer.

Sensitivity tests and contingency tables suggested that there is more persistence in rainfall in the Rocky Mountain and Great Plains states than in areas where moist advection from oceans is present. Seasonal Z and WC correlate to subsequent seasonal temperature better than straight persistence of seasonal temperature. This was especially true in the continental interior. Karl decided that PDSI is not a good predictor of rainfall, perhaps because PDSI is not very sensitive to real changes in soil moisture. Z and WC show skill as predictors, but are still very sensitive to the method of calculating evapotranspiration.

Soil moisture may have usefulness as a predictor outside the interests of atmospheric science. Serafini and Sud (1982) developed a model for the calculation of agricultural drought inception time as a function of soil moisture and atmospheric conditions. They found that for average July soil moisture conditions, drought inception time for North Africa, the Middle East and a large part of western North America was less than 10 days.

Correlation of precipitation with antecedent soil moisture derived from a ground water balance model was computed by Fennessy and Sud (1983) for a 40 year period over the United States. Soil moisture was computed using Thornthwaite's formulation for evaporation relation as modified by Nappo (1975):

$$\frac{\partial w}{\partial t} = P - (1 - e^{-cw/w_s})E_p$$

where P is precipitation, w and w_s are actual and maximum available soil water content, E_p is potential evapotranspiration. The correlation between soil moisture and subsequent precipitation appeared strongest in the western Great Plains in late summer. The high correlation in this drought prone region suggested that a feedback mechanism may be partially responsible for the maintenance of drought.

2.3 Measuring soil moisture

Given its apparent importance, soil moisture has been infrequently and incompletely measured. An exception to this situation can be found in the Soviet Union, where soil moisture has been routinely measured at hundreds of stations in agricultural areas since the 1930s, and over natural surfaces since 1967 (Vinnikov and Yeserkepova, 1991). The method currently used involves removing core samples to a depth of 1.0 to 1.5 m in 10cm segments. Each segment is weighed before and after drying to determine the mass of water contained.

Where widespread direct measurements are not available, soil moisture is often computed from a water-balance relationship. Rasmusson (1968) computed a hydrological budget for North America which included both atmospheric and land branches of the water cycle. He examined North American surface and radiosonde data from 1 May 1958 to 30 April 1963. Surface and subsurface storage change were computed as:

$$\langle \overline{\frac{\partial S}{\partial t}} \rangle = \langle \overline{P - ET} \rangle - \langle \overline{RO} \rangle$$

with evapotranspiration estimated by the methods of Thornthwaite (1964) or Budyko (1963). In the above equation angled brackets indicate spatial average, and overbars indicate time average. Using the relation for change in precipitable water:

$$-\frac{\partial W}{\partial t} = \nabla \cdot Q + (P - ET)$$

where $\nabla \cdot \bar{Q}$ is the water vapor flux divergence, change in soil moisture can also be computed without estimating evaporation using the following balance:

$$\left\langle \frac{\overline{\partial S}}{\partial t} \right\rangle = \langle \nabla \cdot \overline{Q} \rangle + \left\langle \frac{\overline{\partial W}}{\partial t} \right\rangle - \langle RO \rangle$$

where moisture flux and change in precipitable water can be computed from radiosonde observations, and runoff computed from river flow. This vapor balance method gives an annual oscillation in soil moisture two to three times smaller than methods which rely on evaporation estimates. Rasmusson concluded that computed soil moisture is very sensitive to the treatment of evapotranspiration, but that over large areas, estimates can be considered good.

There is promise that ground moisture may also be measurable from space. Wetzel et al. (1984) attempted to deduce soil moisture by satellite using GOES infrared data. They had some success inferring soil moisture from mid-morning rate of surface temperature change with respect to absorbed solar radiation. The method was not applicable when cloud cover was present, and worked best in dry or marginal agricultural regimes. Microwave brightness temperatures, as measured by satellite, can be interpreted to yield soil moisture information in areas of sparse vegetation with greater reliability (Sellers et al., 1990). Goward (1989) has shown that surface reflectance in the visible/near infrared range can be used to infer surface soil wetness for some soil types. Again, this method is only effective where vegetation cover is scant.

3. Modeling Studies

The connection between soil moisture and the atmosphere is through evaporation. It is evaporation which is actually important to circulation. But the connection between soil moisture and evaporation is not completely understood, and often not well simulated. Soil properties, vegetation, and atmospheric conditions all affect evaporation and need to be considered for any complete simulation. Yet current state-of-the-art models rely on parameterizations or simple algorithms to represent this complex process.

Manabe et al. (1965) was the first to incorporate surface hydrology into a general circulation model (GCM), but the land and ocean surfaces were treated as completely wet with no heat capacity. Later, Manabe (1969) modified the model to predict soil moisture and snow cover.

This allowed the land surface to provide actual feedback to the atmosphere by responding to precipitation and surface heating, and altering fluxes to the atmosphere accordingly.

For many years, the state-of-the-art treatment of soil moisture was the "bucket" model of Holloway and Manabe (1971), where the land surface is treated as a grid of reservoirs which are filled with precipitation, emptied by evaporation, and may overflow to produce runoff. Soil moisture availability drops as the surface dries, making evaporation of the remaining moisture — increasingly difficult. This availability was derived by Miyakoda et al. (1979), and is incorporated with varying structures in numerous studies (see Mintz, 1984 for a representative review).

Attempts to improve upon the bucket model have taken several forms. Variations which include two or more soil layers have been developed (Deardorff, 1977; Hansen et al., 1983), and marked improvements to the implementation of the scheme have been realized (Milly, 1992). Meanwhile, some researchers have developed complex surface models which include realistic representation and distribution of vegetation, and its effects on fluxes of heat, moisture and momentum (Rind, 1984; Dickinson et al., 1986; Sellers et al., 1986). Others have pursued statistical approaches which fall between the two in complexity (Entekhabi and Eagleson, 1989; Noilhan and Planton, 1989).

3.1 Sensitivity studies

There have been numerous sensitivity studies of soil moisture using numerical models. In the interest of brevity, only a few of these studies will be reviewed here.

3.1.1 Computing soil moisture

The actual treatment of soil moisture in a GCM can take many forms, and the various forms can be compared. Meehl (1984) examined the effects of specified constant soil moisture versus computed soil moisture. Computed soil moisture led to more realistic seasonal variation in precipitation, particularly in the tropics. Hunt (1985) compared several interactive methods of parameterizing soil hydrology: the simple bucket model of Holloway and Manabe (1971), the two-layer formulation of Hansen et al. (1983), and the two-layer method of Deardorff (1977) which has an extremely shallow (5 mm) upper layer. Deardorff's

formulation was found to give the most realistic results. Meehl and Washington (1988) compared the soil moisture sensitivity of two different GCMs with the same bucket soil representation. It was found that the soil moisture climatologies of the National Center for Atmospheric Research (NCAR) model and the model of the Geophysical Fluid Dynamics Laboratory (GFDL) were quite different, due to the differences between the models in the treatment of radiation and other surface properties.

3.1.2 Simple models

Walker & Rowntree (1977) used a tropical channel model to examine soil moisture effects in sub-Saharan Africa. The channel spanned 36°N to 16°S, and ran from 0°E to 32°E with cyclic boundary conditions on the east and west ends. The soil was represented by a 15 cm bucket model. The simplified surface consisted of land north of 6°N, ocean south. In Case 1 the soil moisture was initially set to 0 cm in the Sahara (14°N-32°N), and 10 cm elsewhere. In Case 2 soil moisture was initially 10 cm everywhere. In case 1, the dry soil region remained dry; in the Sahel region (6°-14°N), evaporation exceeded precipitation by 1.4 mm d^{-1} — a drought situation. In Case 2 the Sahara region gradually dried out, but precipitation spread into the region from the Sahel, which became wetter. After about day 12 precipitation equalled evaporation. The two cases reached different quasi-steady states which reflected the persistence of initial soil moisture. This highly simplified system became intransitive.

Gutman (1984) used a zonally averaged steady-state hemispheric annual-mean model to look at fixed and computed moisture, which he said was analogous to running without and with biofeedback respectively. In specific latitude bands boundary conditions were chosen to simulate desertification, deforestation, and irrigation. The boundary anomalies were held constant in the specified regions in both fixed and biofeedback runs. He found that desertification and deforestation experiments gave similar responses: reduced absorbed radiation, evapotranspiration and precipitation with a concomitant increase in adjacent areas. Irrigation had the opposite effect. Biofeedback produced changes in latitude belts adjacent to anomalies which were of the same order of magnitude as the changes produced by the anomalies themselves. Gutman concluded that biofeedback does not change sign of precipitation response in regions adjacent to anomalies, but can either amplify or moderate the change. In this simple model, perturbations do not modify climate enough to allow continued persistence. A non-stationary model would be needed to study evolution of changes.

3.1.3 GCMs with global anomalies

Shukla and Mintz (1982) performed two summer integrations with extreme soil moisture conditions - one with perpetually saturated ground and one with perpetually dry ground. Significant differences in the global patterns of surface pressure, surface temperature and precipitation were found. As shown in Figure 2, precipitation was greatly reduced and surface temperatures increased as much as 30°C in the dry soil case as compared to the wet soil case. The only region where precipitation was enhanced was over the monsoon region of southern Asia.

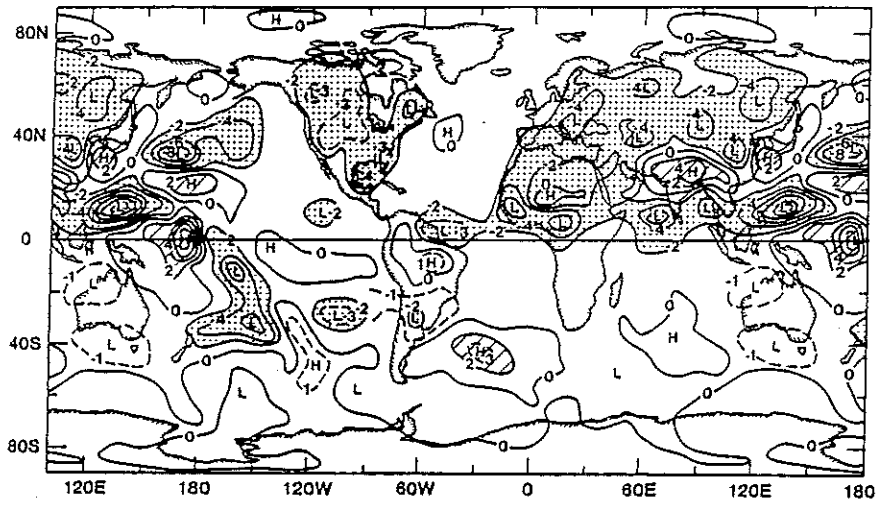

Figure 2. Difference in precipitation for wet minus dry soil cases (Shukla and Mintz, 1982). Units are mm d^{-1}.

Suarez & Arakawa repeated this experiment with the UCLA GCM (Mintz, 1984). The two cases were again dry and saturated soil, and day 16-45 averages were examined. They found that in the wet case, land surface evapotranspiration was 35% higher than that computed by Shukla and Mintz. Precipitation was nearly equal to evaporation over land. In the dry case, almost no rain fell over the continents, except over central Africa. Moisture convergence existed over some land areas but did not produce rain. Mintz (1984) conjectured that discrepancies with Shukla and Mintz were due to model differences, especially in the

parameterizations of the planetary boundary layer and clouds. The UCLA model had fewer clouds, so surface radiation was stronger and evaporation was larger. The lack of precipitation in areas of moisture convergence was apparently due to some of the moisture in the PBL over land being transferred to the free atmosphere by mixing from an unusually strong diurnal cycle over dry land in the UCLA model.

Carson and Sangster (1981) performed GCM experiments with globally saturated and dry initial soil moisture conditions in a bucket model. Evidence of the initial anomalies was still visible in the day 21-50 average precipitation, and some areas still reflected the initial anomalies after 200 days.

3.1.4 GCMs with regional anomalies

There have been numerous experiments with regional scale anomalies. Yeh et al. (1984), using idealized land-sea distributions, found that wet soil moisture anomalies enhanced precipitation only in the mid-latitudes (which were already rainy). However, the anomalies were least persistent in the tropics. This first result has also been found for anomalies over Europe (Rowntree and Bolton 1978; 1983). Simulations with North American soil moisture anomalies (Rind, 1992; Oglesby and Erickson, 1989) imply that droughts can be intensified or prolonged by locally low soil moisture, especially in the interior of the continent. Sud and Smith (1985) found that reduced soil moisture over India seemed to have no effect on precipitation. Experiments in other subtropical areas give mixed results (Sud et al., 1982; Sud and Fennessy, 1984; Kitoh et al., 1988) implying that regional circulation patterns may overwhelm the forcing of the atmosphere by soil moisture anomalies in some areas.

3.1.5 Persistence of anomalies

Delworth and Manabe (1988) used a low-resolution GCM to investigate the character of the persistence of soil moisture anomalies. The GCM used a 15 cm bucket model of hydrology and integrations were carried out for 50 years. Monthly averages from the integration were subtracted from the global fields of soil moisture and precipitation so that only a 50 year record of anomalies remained. It was found that the spectra of precipitation anomalies were nearly white at all latitudes, while the soil moisture spectra were red, with most of the power at very long periods. The redness of the spectra increased from equator to pole. Thus, the

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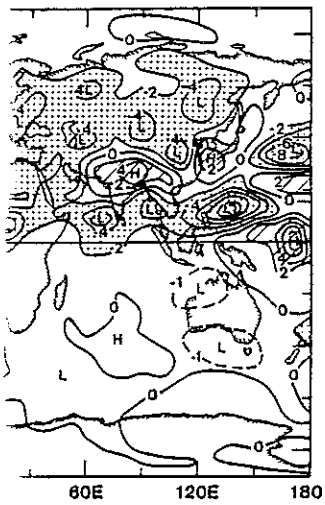

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relationship between soil moisture and precipitation is very closely approximated by a first-order Markov process with precipitation as the forcing, and potential evaporation providing the damping. For their model, half of the variance in soil moisture was at periods greater than 7.5 months in the tropics and subtropics (3°S-31°N), over 12 months in the mid-latitudes (31°-54°N), and 20 months at high latitudes (54°-76°N). However, the model lacked a diurnal cycle, and some important sources of feedback which may alter the persistence of soil moisture anomalies, such as interactive cloudiness and the seasonal variation of potential transpiration, which is caused by the annual cycle of vegetation.

3.1.6 Biosphere models

Sensitivity studies, which have attempted to measure GCM response to the inclusion of vegetation parameterizations, have given tangible if not systematic results. Sato et al. (1989) found that coupling of the Simple Biosphere (SiB) model to a GCM corrected evaporation errors of bucket hydrology. Henderson-Sellers et al. (1990) found that their Biosphere Atmosphere Transfer Scheme (BATS) also seemed to reduce some of the gross errors of the bucket model. In tests of the sensitivity of BATS to various soil characteristics (Wilson et al., 1987), sensitivity to soil texture and upper soil layer depth was found to be high. Although biosphere models depict the process of evapotranspiration more realistically than simpler schemes, they do not necessarily improve on simple schemes in all situations. Also, the vegetation parameterizations in current biosphere models do not react to anomalies or trends in climate, and cannot simulate interannual variability in vegetation cover or vigor.

3.2 Prediction

The overwhelming majority of modeling studies have focused on the sensitivity of the atmospheric response to changes in soil moisture or its formulation. However, a few have examined the feasibility of soil moisture as a predictor of short-term climate. Rind (1982) has investigated the predictive capability of spring soil moisture with relation to summer temperatures and precipitation over the United States. In particular, low springtime soil moisture can be looked upon as a precursor to a hot dry summer. GCM integrations show that while precipitation is reduced, evaporation is reduced more. Thus, "E minus P" decreases and the soil may recharge, erasing the dry anomaly. This processes limits predictability to two or three months - more where the prevailing circulation does not advect moisture from an oceanic source.

The mechanism by which soil moisture anomalies may aid development and persistence of drought in a GCM was explored by Oglesby and Erickson (1989). They determined that a reduction in soil moisture leads to increased surface temperature. The lower atmosphere is heated and ridging occurs aloft. Low level moisture advection is a controlling factor in maintenance of the drought. The degree to which the model circulation is in equilibrium with the soil moisture anomaly determines how quickly the anomaly is diminished.

Meehl (1984) found that inclusion of predicted soil moisture increased the accuracy of monsoon simulations in the NCAR Community Climate Model. He attributed this to the role of soil moisture as a positive feedback over inland regions. This result has been confirmed by Fennessy (personal communication) in experiments where initial soil wetness over India is set very high. Rainfall increased over India as compared to a control run. An experiment with reduced initial soil moisture showed less of a change, but also resulted in increased rainfall. In this case, increased surface heating may be increasing convergence over India. This change in circulation concentrates more moisture over India, similar to what was observed in the study of Shukla and Mintz (1982) with zero soil moisture.

Fennessy and Shukla (1992) performed a similar experiment in a GCM with interactive soil moisture and biosphere. Initial soil moisture was set globally to either climatological values or proxy observed soil wetness derived from the analysis-forecast system of the European Centre for Medium-Range Weather Forecasting (ECMWF). An ensemble of seasonal model integrations were initialized from observed atmospheric states on each of the first three days of June in both 1987 (a non-drought year in central North America) and 1988 (a severe drought year). Soil moisture anomalies were highly persistent. Figure 3 shows the 1 June 1988 initial soil wetness anomalies (ECMWF minus climatology), and the seasonal mean (JJA) soil wetness differences between the two simulations. The integrations with ECMWF initial soil moisture produced a reasonable simulation of 1988 North American drought in both precipitation and surface temperature anomalies, as compared to integrations with climatological initial soil moisture. The simulation of 1988 versus 1987 interannual variability is also ameliorated by use of the "observed" initial soil wetness.

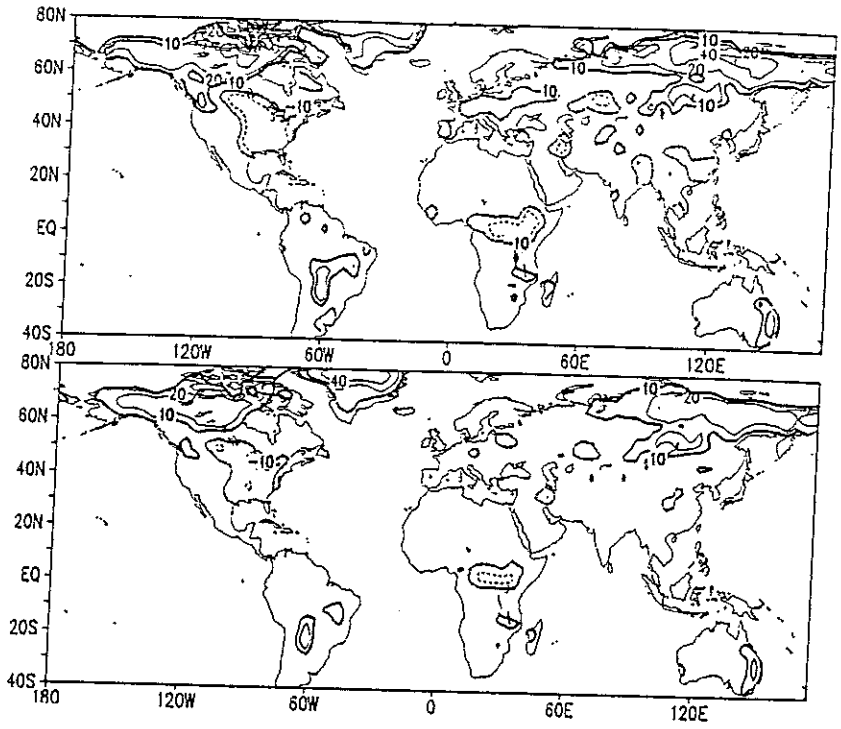
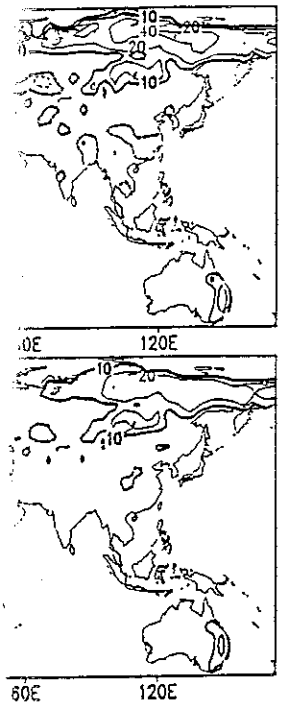


Figure 3: 1 June 1988 soil wetness anomalies (top) and simulated JJA mean soil wetness differences (bottom). Contours are -20, -10, 10, 20, and 40 percent.

In order to use soil moisture as a predictor or indicator of future climate anomalies, accuratesoil moisture measurements must be available to initialize the models, and the models must accurately predict soil moisture. The problems of obtaining comprehensive soil moisture measurements was discussed earlier. Yang et al. (1992) found that relatively small errors in initial soil moisture specification can contribute to sizeable short-term errors in surface air temperature and relative humidity. This is because most of the change in maximum diurnal surface temperature as a function of soil moisture occurs across a narrow range of wetnesses. Yang et al. (1991) have developed a method to correct initial soil moisture based on the computed error of the corresponding surface temperature as predicted by a GCM. Figure 4 shows the mean surface air temperature for the first five days of an ensemble of three summertime forecasts initialized with uncorrected soil moistures. Errors as great as 5°C occur in the semi-arid regions of the western Great Plains. When initial soil moistures are corrected, errors are reduced significantly (Figure 5). Root-mean-square errors also are reduced.



A mean soil wetness differences (bottom).

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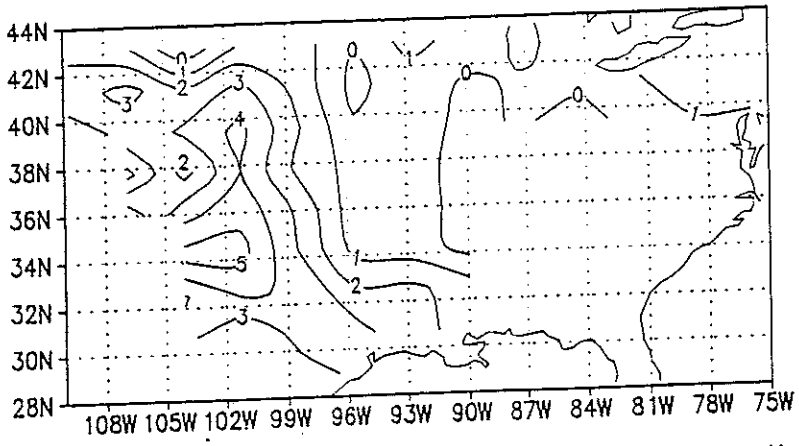

Figure 4: Error in mean surface air temperature for the first five days of forecast with uncorrected soil wetness (Yang et al., 1992). The contour interval is 1°C.

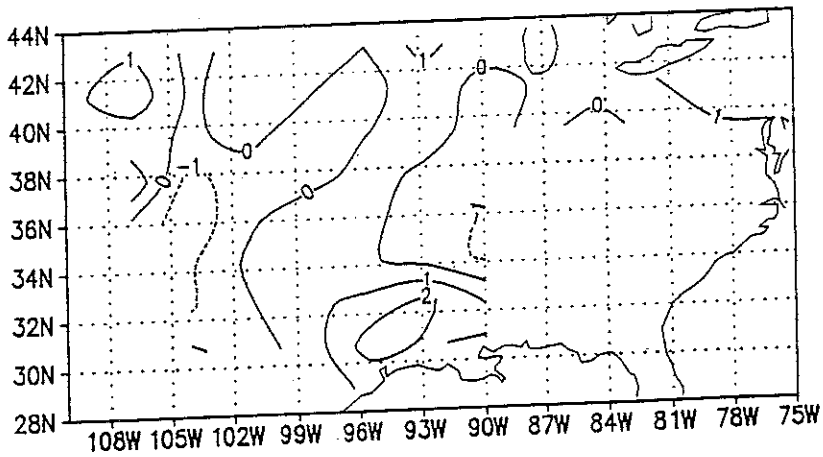

Figure 5: As in Figure 4, except for corrected soil moisture.

Verification of model performance in simulating seasonal soil moisture variations has been conducted by Vinnikov and Yeserkepova (1991). GFDL, Oregon State University (OSU) and United Kingdom Meteorological Office (UKMO) model results were compared to observations over the Soviet Union. All three models use bucket parameterizations, and all grossly underpredicted soil moisture for summer.

3.3 Subgrid variability

Recently there has been a great deal of interest in the role of subgrid scale variability of surface moisture and vegetation in modeling. A number of ways of addressing the problem of representing small-scale variety in large grid boxes have been developed. In the SiB surface model, Dorman and Sellers (1989) use plurality to determine the vegetation and soil type for an entire grid box. Abramopoulos et al. (1988) use area-weighted means of soil and vegetation parameters in a somewhat simpler surface representation. A gaussian distribution of subgrid variations of soil moisture and evaporation is used by Wetzel and Chang (1988) to represent the spottiness of precipitation in wetting a large grid box. Entekhabi and Eagleson (1989) use an exponential distribution of precipitation and soil moisture. Koster and Suarez (1991) have developed a mosaic treatment by which fluxes are computed for each vegetation and soil type found within a given grid box using identical upper boundary conditions. The results from each "tile" of the mosaic are then area-weighted and averaged together to supply a single set of boundary flux values to the atmospheric model.

4. Summary and Conclusions

Soil moisture directly affects the partitioning of energy at the surface between latent and sensible heating. Where soil moisture is high, evaporation will predominate, adding to atmospheric moisture content. Where it is low, the land surface will warm under the influence of radiational heating. Thus, it is not soil moisture which directly affects the atmosphere, but latent and sensible heating which is modulated by soil moisture.

Soil moisture is difficult to quantify in terms of its affect on the atmosphere. Many other factors, such as vegetation, soil characteristics and ambient conditions alter the transfer of moisture from soil to the air. Also, soil moisture is very difficult to measure directly due to its heterogeneity at small scales. There is observational evidence that soil moisture correlates to future rainfall and temperature in certain instances. These correlations seem to be valid only out to three months, and are highly dependent on season and location.

Modeling studies show that strong perturbations in soil moisture on global or regional scales

can affect atmospheric circulation, and persist for several months. However, the role of soil moisture in generating long term climate variability is not well understood. There has been no observational evidence that soil moisture anomalies in an otherwise unaltered surface can persist for time scales beyond a year, nor can they affect interannual climate. Yet changes in surface vegetation such as deforestation (Dickinson and Henderson-Sellers, 1988; Lean and Warrilow, 1989; Nobre et al., 1991) and desertification (Xue and Shukla, 1992) can change the climatological values of soil moisture, as well as surface roughness and albedo. These changes may then alter climate in significant ways. This may be the most important and ominous manifestation of soil moisture repercussions on climate.

REFERENCES

Abramopoulos F, Rosenzweig C and Choudhury B (1988) Improved ground hydrology calculations for global climate models (GCMs): Soil water movement and evapotranspiration. *J Climate* 1:921-941

Anthes R (1984) Enhancement of convective precipitation by mesoscale variations in vegetative covering in semiarid regions. *J Climate Appl Meteor* 23:541-554

Barnston AG and Schickedanz PT (1984) The effect of irrigation on warm season precipitation in the southern Great Plains. *J Climate Appl Meteor* 23:865-888

Budyko MI (1963) *Atlas Teplovogo Balansa Zemnogo Shara*, U.S.S.R. Glavnaya Geofizicheskaya Observatorya, Moscow, 69pp

Carson DJ and AB Sangster (1981) The influence of land-surface albedo and soil moisture content on general circulation model simulations. *Research Activities in Atmospheric and Ocean Modelling*, I D Rutherford, ed. GARP/WCRP Report No. 2, 5.14-5.21

Deardorff JW (1977) A parameterization of ground-surface moisture content for use in atmospheric prediction models. *J Appl Met* 16:1182-1185

Dickinson RE, Henderson-Sellers A, Kennedy PJ and Wilson MF (1986) Biosphere-atmosphere transfer scheme (BATS) for the NCAR Community Climate Model. National Center for Atmospheric Research, Boulder, CO. Tech Note/TN275+STR

Dickinson RE and Henderson-Sellers A (1988) Modelling tropical deforestation: A study of GCM land-surface parametrizations. *Quart J Roy Meteor Soc* 114:439-462

Dickinson RE and Kennedy PJ (1991) Climate effects of Amazon deforestation revisited. *Proc. of the AGU-MSA 1991 Spring Meeting*, Suppliment to Trans Amer Geophys Union American Geophysical Union Washington 82

Dorman JL and Sellers PJ (1989) A global climatology of albedo, roughness length and stomatal resistance for atmospheric general circulation models as represented by the simple biosphere model (SiB). *J Appl Meteor* 28:834-855

Eagleson PS (1991) Hydrologic science: A distinct geoscience. *Rev Geophys* 29:237-248

Entekhabi D and Eagleson PS (1989) Land surface hydrology parameterization for atmospheric general circulation models including subgrid scale spatial variability. *J Climate* 2:816-831

Fennessy MJ and Sud YC (1983) A Study of the Influence of Soil Moisture on Future Precipitation. *NASA Tech. Memo 85042* Goddard Space Flight Center, Greenbelt MD 123p

Fennessy MJ and Shukla J (1992) Influence of initial soil wetness on prediction of global atmospheric circulation and rainfall. J Climate (submitted)

Goward SN (1989) Satellite Bioclimatology. *J Climate* 2:710-720

Gutman G (1984) Numerical experiments on land surface alterations with a zonal model allowing for interaction between the geobotanic state and climate. *J Atmos Res* 41:2679-2685

Hansen J, Russell G, Rind D, Stone S, Lacis A, Lebedeff S, Ruedy R and Travis L (1983) Efficient three-dimensional global models for climate studies: Models I and II. *Mon Wea Rev* 111:609-662

Henderson-Sellers A, Pitman AJ and Dickinson RE (1990) Sensitivity of the surface hydrology to the complexity of the land-surface parameterization scheme employed. *Atmósfera* 3:183-201

Holloway JL and Manabe S (1971) Simulation of climate by a global general circulation model. I. Hydrologic cycle and heat balance. *Mon Wea Rev* 99:335-370

Hunt BG (1985) A modeling study of some aspects of soil hydrology relevant to climatic modelling. *Quart. J Roy Meteor Soc* 111:1071-1085

Karl TR (1983) Some spatial characteristics of drought duration in the United States. *J Climate Appl Meteor* 22:1356-1366

Karl TR (1986) The relationship of soil moisture parameterizations to subsequent seasonal and monthly mean temperature in the United States. *Mon Wea Rev* 114:675-686

Kitoh A, Yamakazi K and Tokioka T (1988) Influence of soil moisture and surface albedo changes over the African tropical rain forest on summer climate investigated with the MRI-GCM-I. *J Meteor Soc Japan* 66 65-86

Koster RD and Suarez MJ (1992) Modeling the land surface boundary in climate models as a composite of independent vegetation stands. *J Geophys Res* 97:2697-2716

Lean J and Warrilow DA (1989) Simulation of the regional climatic impact of Amazon deforestation. *Nature* 342:411-413

Manabe S (1969) Climate and the circulation, I. The atmospheric circulation and the hydrology of the earth's surface. *Mon Wea Rev* 97:739-774

Manabe S, Smagorinsky J and Strickler RF (1965) Simulated climatology of a general circulation model with a hydrological cycle. *Mon Wea Rev* 93:769-798

Meehl GA (1984) Soil moisture, a simple mixed layer ocean and the southern hemisphere semiannual oscillation in the NCAR Community Climate Model. *Studies in Climate*, H. van Loon, ed., NCAR/TN-227+STR, National Center for Atmospheric Research, NTIS No. PB84 196385:115-150

Meehl GA and Washington WM (1988) A comparison of soil-moisture sensitivity in two global climate models. *J Atmos Sci* 45:1476-1492

Milly PCD (1992) Potential evapotranspiration and soil moisture in general circulation models. *J Climate* 5:209-226

Mintz Y (1984) The sensitivity of numerically simulated climates to land-surface boundary conditions. *The Global Climate*, J. Houghton, ed., Cambridge University Press, 79-105

Miyakoda K, Hembree GD and Strickler RF (1979) Cumulative results of extended forecast experiments. II: Model performance for summer cases. *Mon Wea Rev* 107:395-420

Namias J (1959) Persistence of mid-troposheric circulations between adjacent months and seasons. *The Atmosphere and the Sea in Motion*, B. Bolin, ed., Rossby Memorial Volume 240-248

Namias J (1960) Factors in the initiation, perpetuation and termination of drought. *[Extract of publication No. 51*, I.A.S.H. Commision of Surface Waters 81-94

Namias J (1989) Cold waters and hot summers. *Nature* 338:15-16

Nappo CJ (1975) Parameter of surface moisture and evaporation rate in a planetary boundary layer model. J Appl Meteor 14:289-296

Nobre CA, Sellers PJ and Shukla J (1991) Amazon deforestation and regional climate change. *J Climate* 4:957-988

Noilhan J and Planton S (1989) A simple parameterization of land surface processes for meteorological models. *Mon Wea Rev* 117:536-549

Oglesby RJ and Erickson III DJ (1989) Soil moisture and persistence of North American drought. *J Climate* 1:1362-1380

Palmer WC (1965) Meteorological drought. Res Pap 45 U.S. Weather Bureau, 58pp

Rasmusson EM (1968) Atmospheric water vapor transport and the water balance of North America II. Large-scale water balance investigations. *Mon Wea Rev* 96:720-734

Rind D (1982) The influence of ground moisture conditions in North America on summer climate as modeled in the GISS GCM. *Mon Wea Rev* 110:1487-1494

Rind D (1984) The influence of vegetation on the hydrologic cycle in a global climate model. *Climate Processes and Climate Sensitivity*, J. E. Hansen and T. Takahashi, eds., Geophysical Monograph 26 Maurice Ewing 5:73-91

Rowntree PR and Bolton JA (1978) Experiments with soil moisture anomalies over Europe. *The GARP Programme on Numerical Experimentation:* Research Activities in Atmospheric and Ocean Modelling, R. Asselin ed., Report No. 18 WMO/ICSU Geneva 63

Rowntree PR and Bolton JA (1983) Effects of soil moisture anomalies over Europe in summer. *Variations in the Global Water Budget*, A. Street-Perrott et al., eds, D Reidel 447-462

Salati E and Nobre CA (1991) Possible climatic impacts of tropical deforestation. *Climatic Change* (in press)

Sato N, Sellers PJ Randall DA Schneider EK Shukla J Kinter III JL Hou Y-T and Albertazzi E (1989) Effects of implimenting the Simple Biosphere model in a general circulation model. *J Atmos Sci* 46:2757-2782

Sellers PJ, Mintz Y Sud YC and Dalcher A (1986) A simple biosphere model (SiB) for use within general circulation models. *J Atmos Sci* 43:505-531

Sellers PJ, Rasool SI and Bolle H-J (1990) A review of satellite data algorithms for studies of the land surface. *Bull Amer Meteor Soc* 71:1429-1447

Serafini YG and Sud YC (1982) A model for the inception time of an agricultural drought. *Research Activities in Atmospheric and Ocean Modelling*, ID Rutherford, ed. GARP/WCRP Report No 3:5.38-5.43

Shukla J and Mintz Y (1982) Influence of land-surface evapotranspiration on the earth's climate. *Science* 215:1498-1501

Sud YC Fennessy M and Shukla J (1982) Influence of evapotranspiration anomalies in semi-arid regions on July circulation. *Research Activities in Atmospheric and Ocean Modelling*. ID Rutherford, ed. GARP/WCRP Report No 3:5.30-5.33

Sud YC and Fennessey MJ (1984) Influence of evaporation in semi-arid regions on the July circulation: A numerical study. *J Climatology* 4:383-398

Sud YC and Smith WE (1985) Influence of local land-surface processes on the Indian monsoon: A numerical study. *J Climate Appl Meteor* 24:1015-1036

Thornthwaite CW and Associates (1964) Average climatic water balance data of the continents: Part VII. United States. *Publications in Climatology*, Laboratory of Climatology, Centerton, NJ 17:419-615

Vinnikov K Ya and Yeserkepova IB (1991) Soil moisture, empirical data and model results. *J Climate* 4:66-79

Walker J and Rowntree PR (1977) The effect of soil moisture on circulation and rainfall in a tropical model. *Quart J Roy Meteor Soc* 103:29-46

Walsh JW, Jasperson H and Ross B (1985) Influences of snow cover and soil moisture on monthly air temperature. *Mon Wea Rev* 113:756-768

Wetzel PJ and Chang J-T (1988) Evapotranspiration from nonuniform surfaces: A first approach for short-term numerical weather prediction. *Mon Wea Rev* 116:600-621

Wetzel PJ, Atlas D and Woodward RH (1984) Determining soil moisture from geosynchronous satellite infrared data: A feasibility study. *J Climate Appl Meteor* 23:375-391

Wilson MF, Henderson-Sellers A, Dickinson RE and Kennedy PJ (1987) Sensitivity of the Biosphere-Atmosphere Transfer Scheme (BATS) to the inclusion of variable soil characteristics. *J Climate Appl Meteor* 26:341-362

Xue Y and Shukla J (1992) A numerical experiment to study the influence of changes in the land surface properties on Sahel climate: Part 1 - Deforestation. *J Climate* (submitted)

Yang R, Fennessy MJ and Shukla J (1992) The influence of the variation of initial soil wetness on the surface weather and atmospheric circulation. *Mon Wea Rev* (submitted)

Yeh T-C, Wetherald RT and Manabe S (1984) The effect of soil moisture on the short-term climate and hydrology change - A numerical experiment. *Mon Wea Rev* 112:474-490