

Impact of Initial Soil Wetness on Seasonal Atmospheric Prediction

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(Manuscript received 29 January 1998, in final form 7 October 1998)

ABSTRACT

This study investigates the importance of initial soil wetness in seasonal predictions with dynamical models. Two experiments are performed, each consisting of two ensembles of global climate model integrations initialized from early June observed atmospheric states. In each experiment the only difference between the two ensembles is that they are initialized with a different soil wetness. In the first experiment both ensembles are initialized from 1988 observed atmospheric states and use observed 1988 SST; one ensemble is initialized with seasonally varying climatological soil wetness, and the other is initialized with proxy 1988 soil wetness derived from the European Centre for Medium-Range Weather Forecasts analysis–forecast system. In the second experiment the two ensembles are initialized from observed atmospheric states and use observed SST for five different years, and each ensemble is initialized with a different climatological soil wetness. After initialization, a coupled atmosphere–biosphere model determines the evolution of the soil wetness fields in all the integrations.

The experiments are analyzed to determine the impact of the initial soil wetness differences. In contrast to several previous studies in which initial soil wetness was prescribed arbitrarily, a somewhat more realistic soil wetness impact is analyzed by comparing integrations initialized with climatological soil wetness to integrations initialized with soil wetness derived from the output of an operational analysis–forecast model. The initial soil wetness impact is found to be largely local and is largest on near-surface fields, in agreement with previous results. Significant impacts were found in several tropical and extratropical regions in both experiments. Almost all the regions that had significant increases (decreases) in initial soil wetness had significant increases (decreases) in seasonal mean evaporation and significant decreases (increases) in seasonal mean surface air temperature. Half of the regions had significant increases (decreases) in seasonal mean precipitation in response to increased (decreased) initial soil wetness, though the response of the precipitation was more variable and was highly dependent on the response of the moisture flux convergence to the initial soil wetness anomaly. In order for an initial soil wetness difference to force a significant seasonal mean precipitation difference in a region, it must effectively alter the mean convective stability of the region and thereby the mean convective precipitation.

The strength of the impact of initial soil wetness differences, as well as the nature of the impact on precipitation and other atmospheric fields, depends on several factors. These factors include the areal extent and magnitude of the initial soil wetness difference, the persistence of the soil wetness difference, the strength of the solar forcing, the availability of nearby moisture sources, and the strength of the regional dynamical circulation. The results suggest that seasonal atmospheric prediction could be enhanced by using a realistic initial state of soil wetness.

1. Introduction

Over the past 20 years there has been a large number of observational and modeling studies investigating the role of water stored in the soil in determining the mean climate and its interannual variability; a recent review is given by Dirmeyer and Shukla (1993). Here we shall give only a brief summary of several modeling studies.

With a zonally symmetric tropical model, Walker and Rowntree (1977) demonstrated that initially imposed

soil wetness anomalies could persist and have a significant impact on 20-day simulations. Miyakoda et al. (1979) found that specifying soil wetness from past precipitation data yielded improved medium-range general circulation model (GCM) forecasts of precipitation and evaporation. Shukla and Mintz (1982) showed that global fields of rainfall, temperature, and motion strongly depend on land surface evapotranspiration in an extremely hypothetical experiment with either zero or maximum evapotranspiration over all land surfaces in a global GCM. Rind (1982) reduced the U.S. ground moisture to one-quarter of its value in control integrations and found a significant impact on the boreal summer climate. Sud and Fennessy (1984) found that prescribing zero evapotranspiration even in four semiarid regions substantially affected the local climate simulated by a global GCM. In a review of 11 land surface sen-

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sitivity studies Mintz (1984) concluded that soil moisture had a stronger impact on model-simulated climates than did albedo or surface roughness.

Delworth and Manabe (1988) found that interactive soil moisture (as opposed to prescribed soil moisture) substantially increased summer surface air temperature variability in GCM integrations. Delworth and Manabe (1989) showed that soil moisture fluctuations result in significant variations of near-surface atmospheric humidity and temperature. They also concluded that due to the seasonal and interannual timescales of soil moisture anomalies, they can increase the persistence of near-surface atmospheric humidity and temperature.

A number of modeling studies have emphasized the importance of soil moisture anomalies in droughts. Oglesby and Erickson (1989) demonstrated the important role of soil moisture in prolonging and/or amplifying North American summer drought. Oglesby (1991) found that the timing of spring soil moisture anomalies was crucial in determining their impact on the succeeding summer climate. Branković et al. (1990) found that the veracity of 1988 summer GCM simulations over North America was highly dependent on using real-time initial soil wetness from the European Centre for Medium-Range Weather Forecasts (ECMWF) analysis–forecast system in lieu of climatological initial soil wetness. Atlas et al. (1993) found that the use of proxy 1988 soil moisture anomalies resulted in precipitation anomalies and significant surface temperature anomalies over the Great Plains region. Fennessy et al. (1994) found that proxy 1988 soil wetness contributed to the 1988 U.S. drought precipitation anomalies during June and the surface temperature anomalies during June, July, and August in simulations with the Center for Ocean–Land–Atmosphere Studies (COLA) GCM.

Soil moisture anomalies have also been linked to the 1993 U.S. summer floods. Betts et al. (1994) found that the July 1993 precipitation in the central United States was doubled in a GCM integration initialized with saturated soil as opposed to one initiated with soil dry enough that the surface resistance to evaporation was four times the unstressed value. Paegle et al. (1996) concluded that accurate surface evaporation in the Great Plains region is necessary for accurate simulations of dynamic support for rainfall in a regional model study.

Soil moisture has also been demonstrated to be important in idealized climate model studies. Cook and Gnanadesikan (1991) found that saturated versus dry land surfaces had a dramatic impact on the land–sea temperature contrast and precipitation distribution in a GCM study with idealized boundary conditions. Dirmeyer (1994) investigated the effect of soil moisture on drought in midlatitudes in a coupled atmosphere–biosphere GCM study with an idealized land–sea distribution. He found that low spring soil moisture could lead to summer drought conditions.

The focus of the current study is on the impact of initial soil wetness on seasonal atmospheric prediction.

We perform pairs of integrations that start from observed atmospheric initial conditions and use observed SST, the only difference being the initial soil wetness. We recognize that there are no global measurements of soil wetness, and in fact there is considerable uncertainty in defining the soil moisture for any grid box. Part of the difficulty arises because of heterogeneity of the land surface, with different soil and vegetation types within a model grid box. Topographic undulations of the land surface cause additional complexity in defining soil wetness. Nonetheless, GCMs are quite sensitive to the numerical value of soil wetness, and GCMs with current parameterizations require that numerical values be assigned at each land grid point. The different initial soil wetness fields used here have been widely used by the modeling community. Though they are not actual observations of soil wetness, they are realistic in that they have been used in models in lieu of observed soil wetness. In this sense, the differences between them and their impact on seasonal prediction may be viewed as realistic, rather than idealized or extreme, as has been the case in almost all of the previously mentioned studies. Thus, it is important to determine if the differences between these soil wetness fields actually used to initialize prediction models have a significant impact on seasonal predictions with such models. After initialization, the soil wetness itself is predicted by the coupled atmosphere–biosphere GCM.

In the first experiment, we analyze the impact of using proxy 1988 soil wetness versus climatological soil wetness in ensembles of 1988 integrations. We should emphasize that during the spring and early summer of 1988 there were highly anomalous drought conditions over the central United States, which partially motivated this study. Thus, the results for this region (and perhaps others) may be quite different if another year were chosen. In order to simulate the possible impact of 1988 soil wetness on the 1988 seasonal climate, we use initial atmospheric states from 1988 in all of the integrations; that is, the initial atmospheric states used in each ensemble are not independent. Because it is difficult to define a global observed soil wetness field, we perform integrations initialized with climatological soil wetness calculated by Willmott et al. (1985) and with proxy 1988 soil wetness from the ECMWF analysis–forecast system. Although this experiment is intended to examine the possible impact of observed 1988 soil wetness on 1988 seasonal predictions, it is also possible that some of the simulated impact is due to relative biases in the soil wetness fields, that is, differences between the respective climatologies of the two soil wetness fields used. This possibility is examined in the second experiment.

In the second experiment, we analyze the impact of using two different climatological soil wetness fields to initialize five member ensembles. The five members of these ensembles use SST and initial atmospheric states from five different years and are thus independent. One

ensemble is initialized with the same climatological soil wetness field used in the first experiment. The second ensemble is initialized with a climatology calculated from the soil wetness time series from the ECMWF analysis–forecast system for 1987–93, from which the 1988 soil wetness used in the first experiment was obtained.

Although the design of the two experiments is somewhat different, both address the basic question of the importance of moderate (and hopefully realistic) soil wetness variations in seasonal prediction. The results from the two experiments will be compared in order to strengthen the general conclusions made in this study.

The results will be analyzed to determine the impact of the initial soil wetness and to investigate what factors play a role in determining this impact. The model and details concerning the soil wetness fields used are described in section 2. The results from the proxy 1988 initial soil wetness experiment are presented in section 3. The results from the different climatological initial soil wetness experiments are presented in section 4. A summary is given in section 5.

2. Model and soil wetness initialization

a. Model

The COLA GCM is based on a modified version of the National Centers for Environmental Prediction (NCEP) global spectral model used for medium-range weather forecasting [see Sela (1980) for original NCEP formulation; see Kinter et al. (1988) for the modified version]. The land surface parameterization was changed to the simple biosphere model (SiB) biophysical formulation after Sellers et al. (1986) by Sato et al. (1989a) and was later simplified by Xue et al. (1991). The model used in the first (1988) set of experiments uses Kuo deep convection (Anthes 1977; after Kuo 1965) and Tiedtke (1984) shallow convection and is the same as presented by Fennessy et al. (1994). The more recent model used in the second (climatological soil wetness) set of experiments uses relaxed Arakawa–Schubert convection (Moorthi and Suarez 1992; after Arakawa and Schubert 1974) and Tiedtke (1984) shallow convection after Hogan and Rosmond (1991) and is described by Dewitt (1996).

The COLA GCM is a global spectral model with rhomboidal truncation at zonal wavenumber 40. The model physics calculations are done on a 1.8° lat \times 2.8° long Gaussian grid. The vertical structure of the model is represented by 18 unevenly spaced levels using σ as the vertical coordinate (Phillips 1957). The spacing of the levels is such that greater resolution is obtained near the earth's surface and at the tropopause. In addition to the parameterizations mentioned above, the COLA GCM includes parameterizations of solar radiative heating (Lacis and Hansen 1974), terrestrial radiative heating (Harshvardhan et al. 1987), large-scale condensa-

tion, interactive cloud radiation (Hou 1990; after Slingo 1987), gravity wave drag (Vernekar et al. 1992; after Alpert et al. 1988), and a turbulence closure scheme for subgrid-scale exchanges of heat, momentum, and moisture (Miyakoda and Sirutis 1977; Mellor and Yamada 1982).

In the COLA GCM, each land grid box (approximately 1.8° lat \times 2.8° long) is assigned one of 12 sets of vegetation and soil characteristics, based on the dominant vegetation observed in the grid box (Dorman and Sellers 1989). Included in these characteristics are the depth and porosity of each of three soil layers: the surface layer, the root zone, and the drainage layer. The total depth of the three layers ranges from 49 cm for bare soil (desert) to 350 cm for trees. The porosity ranges from 0.42 to 0.46. The total water-holding capacity of a grid box is the sum of the products of the porosity and soil depth for each layer. The soil wetness of a grid box is defined as the ratio of the total water present to the total water-holding capacity. The total water-holding capacity ranges from 21 cm for bare soil to 147 cm for trees.

b. Soil wetness initialization

Initialization and validation of GCM soil wetness fields has long been recognized as a challenging problem (Sellers et al. 1986, 1989; Sato et al. 1989a,b). Mintz and Serafini (1981) used observed precipitation and surface air temperature in a simple water budget model to produce a global monthly climatology of soil moisture, which has since been widely used in atmospheric models (Mintz and Serafini 1992). It should be possible to initialize GCM integrations with observed soil moisture data obtained through the use of a hybrid methodology incorporating remotely sensed data with sophisticated coupled atmosphere–biosphere models (Sellers 1990; Liston et al. 1993). As this sort of hybrid soil wetness data is not yet available, we study the impact of initializing the COLA GCM with proxy 1988 soil wetness derived from the prognostic soil moisture that is produced by the ECMWF operational analysis–forecast cycle.

The climatological soil wetness used to initialize the control integrations was derived from the monthly climatological soil moisture calculated by Willmott et al. (1985) and is similar to that of Mintz and Serafini (1992), which has been used in many atmospheric models. The proxy 1988 initial soil wetness used in the first set of experiments is derived from the ECMWF analysis–forecast cycle produced operationally for 1 June 1988. This version of the ECMWF model uses three soil layers; the top two layers are prognostic and the bottom layer is updated to time-varying climatological values. Both ECMWF prognostic soil moisture layers, the surface layer with a maximum capacity of 20 mm of liquid water and the deep soil layer with a maximum capacity of 120 mm of liquid water, were used for ini-

tialization. The bottom ECMWF soil moisture layer was not used. Starting in November 1983, the ECMWF forecast model has been integrated in a four-dimensional data-assimilation mode with continuously updated soil moisture, from which a soil moisture time series is available.

Unfortunately, the Willmott et al. climatological soil moisture and the ECMWF soil moisture cannot be used directly by the SiB. The land surface parameterizations (LSPs) used by Willmott et al. and in the ECMWF model are different from the SiB, so the resulting soil moisture fields are different from what would have been calculated by the SiB if exposed to the same atmospheric forcings. Sato et al. (1989b) developed a method to transform soil moisture calculated by other LSPs to be consistent with the SiB. This method was used to transform the Willmott et al. climatological soil moisture. Essentially, the method calculates the time integral of evaporative demand that a GCM grid area would have to be exposed to dry down from saturation to a specified level. This same time integral is then applied to a greatly reduced set of SiB energy balance equations to calculate an equivalent SiB soil moisture level. This procedure is reviewed in Fennessy et al. (1994). Soil wetness is the soil moisture content expressed as a fraction or percent of the maximum liquid water capacity for a given layer. The SiB model carries three prognostic soil wetnesses: that of the surface layer, that of the root zone, and that of the gravitational drainage zone. The procedure outlined by Fennessy et al. (1994) is used to transform the two combined ECMWF prognostic soil moisture layers into a soil wetness that is used to initialize all three SiB prognostic soil wetness layers. It should be noted that the ECMWF soil moisture described here and the ECMWF LSP parameterization described by Fennessy et al. (1994) were in use operationally during 1987–93 and are significantly different from those used more recently by ECMWF.

3. Impact of proxy 1988 soil wetness on seasonal prediction

Two three-member COLA GCM integration ensembles were conducted. A summary of the integration ensemble names, integration initiation dates (all 0000 UTC), integration lengths, SST boundary condition used, and soil wetness initial condition used is given in Table 1.

To study the influence of observed initial soil wetness on seasonal prediction during 1988 we compare a 90-day GCM ensemble initialized in early June 1988 with climatological soil wetness (CON88) to an ensemble differing only in that it was initialized with proxy 1988 soil wetness (SW88). To make the experiment as realistic as possible we use observed SST and observed atmospheric initial conditions from 1988. If we chose independent random initial conditions and SST from different years, the results might be different but also

might less accurately portray the impact of surface forcing during 1988. The integrations were initialized from the NCEP analyses of the observed atmospheric states at 0000 UTC on 1, 2, and 3 June 1988. In all the integrations, observed time-varying SST (Reynolds 1988) was used and the soil wetness was predicted.

The 1 June climatological soil wetness used to initialize CON88 is shown in Fig. 1a. The regions where the magnitude of the SW88 – CON88 difference in initial soil wetness is five or more standard deviations of the local ECMWF soil wetness are shaded in Fig. 1b. The standard deviation used is the mean of the May and June monthly standard deviations calculated from the 1987–93 time series. Dark shading indicates positive differences; light shading indicates negative differences. The larger, spatially coherent regions that meet this criteria are boxed and numbered (1–9) in Fig. 1b and will be analyzed individually. The region numbers, names, areal boundaries, and area-averaged initial soil wetness (ISW) differences (percent of saturation) are given in Table 2. These regions have been chosen based solely on this objective initial soil wetness criteria, rather than from an a posteriori examination of the results. By selecting the study regions in this fashion we hope to identify factors important in determining the impact of significant regional initial soil wetness differences. An examination of global maps of the simulated initial soil wetness impact in the evaporation, precipitation, winds, height, and temperature fields reveals that the impact is local to the regions with significant initial soil wetness anomalies (not shown). The results presented here will all be three-case ensemble means averaged over the exact regions depicted in Fig. 1b and given in Table 2.

The 90-day mean SW88 – CON88 difference in root zone soil wetness (SWR, percent of saturation), evaporation (EVAP, mm day⁻¹), surface temperature (TS, °C), precipitation (P, mm day⁻¹), and vertically integrated moisture flux convergence (VMFC, mm day⁻¹) are given for the nine study regions in Table 3. For convenience the ISW difference (percent of saturation, as in Table 2) is also given in Table 3. Bold type in Table 3 indicates area-averaged differences that were significant at the 95% level, as determined by a *t* test.

The root zone soil wetness variability is a good indicator of the three-layer mean soil wetness variability. The surface soil wetness response is quicker; that of the drainage zone is slower. The 90-day mean root zone soil wetness differences are significant in all nine study regions and are of the same sign as the initial soil wetness differences, suggesting relatively strong persistence of the initial soil wetness differences (Table 3). This persistence is more readily seen in the area-averaged daily time series of the CON88 and SW88 root zone soil wetness shown in Fig. 2. In this and the remaining time series shown, the CON88 three-case ensemble is solid and the SW88 three-case ensemble is dotted. The most remarkable feature in Fig. 2 is that the root zone soil wetness differences remain throughout the entire 90

TABLE 1. Integration ensemble names, initial dates, lengths, SST used, and soil wetness initial condition (IC) used in 1988 initial soil wetness experiment.

Ensemble name	Initiation dates	Length (days)	SST	Soil wetness IC
CON88	1, 2, 3 Jun 1988	90	1988 Observed ^a	Climatological ^b
SW88	1, 2, 3 Jun 1988	90	1988 Observed ^a	1988 ECMWF ^c

^a Reynolds 1988.

^b From Willmott et al. 1985.

^c From 1 Jun 1988 ECMWF analysis–forecast cycle soil moisture.

days in eight of the nine study regions. In region 4, southeast China, the soil wetness difference persists for roughly 60 days. This region had one of the smallest initial soil wetness differences (-11.3%) and one of the smallest areal extents of the nine study regions (Fig. 1b).

The 90-day mean evaporation differences (Table 3, column 4) are also significant in all nine study regions and are of the same sign as the soil wetness differences. However, it is clear that there is no simple relation between the magnitude of the soil wetness differences and the magnitude of the evaporation differences across the nine study regions. The area-averaged daily time series of the CON88 and SW88 evaporation is shown in Fig. 3. The CON88 versus SW88 differences in the evaporation time series (Fig. 3) tend to follow those in the soil wetness time series in four of the study regions (regions 1, 2, 4, and 5) but not in southeast Russia (region 3), south equatorial Africa (region 6), east Australia (region 7), Alaska–Canada (region 8), and Siberia (region 9). Due to their latitude, the solar forcing (not shown) is relatively weak in four of these regions (regions 3, 7, 8, and 9). In east Australia (region 7) the solar forcing is so weak and the surface temperatures are so cold that little evaporation occurs (1 mm day^{-1}) in either ensemble, and there is no response to the soil wetness difference, which is very persistent. In the relatively cold, high-latitude regions (regions 3, 8, and 9) the maximum evaporation differences occur in midsummer, also the time of maximum surface temperatures.

The evaporation in the south equatorial Africa region (region 6) is also insensitive to the persistent soil wetness differences, despite relatively strong solar forcing and warm surface temperatures. This region is very dry, with simulated precipitation and evaporation rates below 1 mm day^{-1} throughout the season (not shown). This region had the smallest initial soil wetness difference of the study regions (7.2%), which resulted in a small but significant evaporation difference (0.1 mm day^{-1}), which was unable to significantly impact the surface temperature or precipitation.

The 90-day mean ensemble surface temperature differences (Table 3, column 5) are opposite in sign to the evaporation and soil wetness differences in all seven study regions that had significant surface temperature differences, indicating a strong tendency for warmer (colder) surface temperatures with reduced (increased)

evaporation. This occurs because a larger (smaller) fraction of the radiation goes toward heating the ground rather than toward evaporating water. The 90-day mean surface temperature differences are small ($<0.5^\circ\text{C}$) and statistically insignificant in regions 6 and 7, which had the smallest 90-day mean evaporation differences (0.2 mm day^{-1} or less).

The 90-day mean precipitation differences (Table 3, column 6) are statistically significant in four regions (regions 1, 3, 8, and 9), where they are of the same sign as the soil wetness and evaporation differences. In regions 3, 8, and 9 they are of approximately the same magnitude as the evaporation differences. In region 1 (eastern United States) the precipitation increase is half the evaporation increase, with the other half accounted for by significantly decreased 90-day mean vertically integrated moisture flux convergence (Table 3, column 7). The 90-day mean vertically integrated moisture flux convergence was significantly decreased by 0.8 mm day^{-1} in region 5 (Bolivia–Brazil), which had a negligible 90-day mean precipitation difference, despite having one of the larger 90-day mean evaporation differences (0.7 mm day^{-1}). Bolivia–Brazil is the only one of the four regions with a 90-day mean evaporation difference greater in magnitude than 0.5 mm day^{-1} that did not have a same-sign, statistically significant 90-day mean difference in precipitation. The 90-day mean total precipitable water differed only negligibly between the CON88 and SW88 ensembles in the nine study regions (not shown). The daily time series of precipitation and vertically integrated moisture flux convergence over each study region (not shown) are similar and highly variable, indicative of the impact of large-scale forcing, which is highly variable and, in general, not directly related to the soil wetness differences.

The 90-day mean sea level pressure was significantly different only in the eastern United States (region 1), where it was reduced by 1.8 hPa in the SW88 ensemble (not shown). The significant seasonal mean evaporation, surface temperature, precipitation, and sea level pressure differences in the eastern United States region are reflective of the impact of the proxy 1988 initial soil wetness on the 1988 U.S. drought. The 1988 U.S. drought signal is also evident in the eastern United States region soil wetness, evaporation, and surface temperature time series (Figs. 2–4).

There was little impact on upper-level fields. The 300-

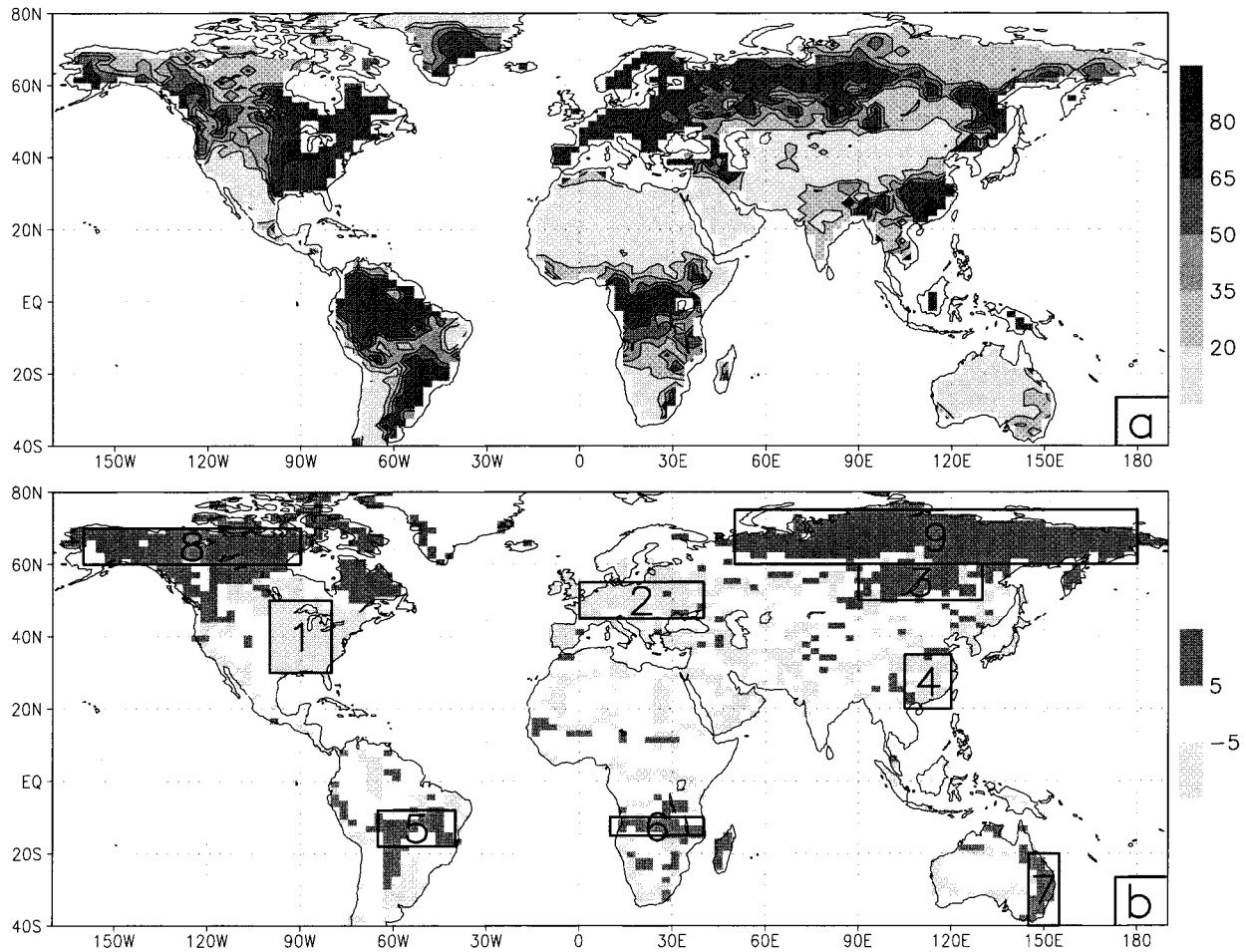


FIG. 1. (a) Soil wetness climatology (% of saturation) for 1 Jun used to initialize CON88 and CL1 ensembles and (b) initial soil wetness difference in SW88 - CON88 ensembles. Regions with differences greater/less than 10% of saturation are shaded dark/light.

mb geopotential heights were significantly different only over Alaska-Canada (region 8), where a reduction by 51 geopotential meters occurred in the SW88 ensemble (not shown). This region had the largest 90-day mean differences in surface temperature (-3°C) and

precipitation (1 mm day^{-1}) and is also one of the largest of the study regions in areal extent.

4. Impact of initial soil wetness from two different climatologies

In the previous section, we examined the impact of initial soil wetness on seasonal predictions made from observed initial conditions for 1988. In this section, we would like to investigate whether the conclusions from the case study of the summer of 1988 also hold for initial soil wetness from two different soil wetness climatologies using an ensemble of integrations with different initial atmospheric states. For this purpose we have used a more recent and significantly improved version of the COLA GCM.

Two sets of seasonal integrations were carried out using initial atmospheric conditions from NMC analyses at 0000 UTC on 1 June 1987, 1988, 1990, 1992, and 1993. In all the integrations, the soil wetness was predicted after initialization, and observed time-varying

TABLE 2. Study regions, extents and area-averaged initial soil wetness differences (% of saturation) for 1988 initial soil wetness experiment.

#	Region	Bounds	ISW anomalies (%)
1	Eastern United States	30°-50°N, 80°-100°W	-29.7
2	Central Europe	45°-55°N, 0°-40°E	-10.4
3	Southeast Russia	50°-60°N, 90°-130°E	14.9
4	Southeast China	20°-35°N, 105°-120°E	-11.3
5	Bolivia-Brazil	8°-18°S, 40°-65°W	13.4
6	South equatorial Africa	10°-15°S, 10°-40°E	7.2
7	East Australia	20°-40°S, 145°-155°E	23.6
8	Alaska-Canada	60°-70°N, 90°-160°W	30.5
9	Siberia	60°-75°N, 50°-180°E	27.9

TABLE 3. Area-averaged initial soil wetness difference (ISW, % of saturation) and 90-day mean SW88 – CON88 difference in root zone soil wetness (SWR, % of saturation), evaporation (EVAP, mm day⁻¹), surface temperature (TS, °C), precipitation (P, mm day⁻¹), and vertically integrated moisture flux convergence (VMFC, mm day⁻¹) for the nine study regions in the 1988 initial soil wetness experiment. Bold type indicates area-averaged differences that were significant at the 95% level, as determined by a *t* test.

#	ISW	SWR	EVAP	TS	P	VMFC
1	-29.7	-14.8	-1.1	2.7	-0.5	0.6
2	-10.4	-6.5	-0.4	0.9	-0.3	0.1
3	14.9	13.3	0.6	-1.4	0.7	0.2
4	-11.3	-4.3	-0.4	1.2	-0.3	0.2
5	13.4	8.2	0.7	-1.2	-0.1	-0.8
6	7.2	5.4	0.1	0.1	0.1	0.0
7	23.6	11.9	0.2	-0.4	0.0	-0.2
8	30.5	20.7	0.9	-3.0	1.0	0.2
9	27.9	15.4	0.4	-1.3	0.3	-0.1

SST (Reynolds and Smith 1994) was used. In one set of integrations, referred to as CL1 (see Table 4), the initial soil wetness on 1 June was derived from the Willmott et al. (1985) climatology; and in the other set, referred to as CL2, the initial soil wetness on 1 June was derived from the ECMWF analysis–forecast cycle soil moisture averaged for 7 yr (1987–93).

To obtain the ECMWF soil wetness climatology, daily values of the two levels of ECMWF prognostic soil moisture for 1987–93 were obtained from the archives at the National Center for Atmospheric Research. The daily values were transformed into SiB-compatible soil wetness by a method similar to that described in section 2b. Monthly means and a monthly climatology were then formed using the data for 1987–93, except the period from August 1989 through March 1990. The ECMWF soil wetness was erroneously low from August 1989 to March 1990 due to an operations error that resulted in the layer-3 soil wetness being retained at August climatological values for the entire period (Č. Branković 1991, personal communication).

The soil wetness for 1 June used to initialize these integrations was obtained by linear interpolation of the monthly climatological values. The soil wetness used to initialize CL1 and CL2 is shown in Figs. 1a and 5a, respectively. The regions where the magnitude of the CL2 – CL1 difference in initial soil wetness is five or more standard deviations of the local ECMWF soil wetness are shaded in Fig. 5b. As in section 3, the standard deviation used is the mean of the May and June monthly standard deviations calculated from the 1987–93 time series. Dark shading indicates positive differences; light shading indicates negative differences. The larger, spatially coherent regions that meet this criteria are boxed and labeled in Fig. 5b and are described in Table 5. It is notable that in most of the regions the initial soil wetness differences are positive, indicating that the climatology used in CL2 is generally wetter than that used in CL1. There are several regions that are in common

with those used in the 1988 experiment (numbered 3, 5, 7, 8, and 9), and the sign and magnitude of the initial soil wetness differences as well the impact in these regions is quite similar to that described in section 3. This implies that the initial soil wetness differences in these five regions are due more to differences in the respective climatologies of the soil wetness used than to anomalous 1988 values. The similarity of the impact in these five regions in the two experiments increases the robustness of the results. In this section we will concentrate on the four new regions labeled A, B, C, and D.

The impact of the different initial climatological soil wetness in each of regions A, B, and C (western United States, Southeast Asia, and southern Africa, respectively) is similar, with significant increases in seasonal mean root zone soil wetness and evaporation and decreases (increases) in seasonal mean surface temperature (precipitation), which were significant in two of these three regions (Table 6). In region D (Saudi Arabia) the small reduction in initial soil wetness (-1.4%) was insufficient to force a significant model response.

Both of the soil wetness climatologies used to initialize ensembles CL1 and CL2 are calculations using atmospheric observations as input. A question of practical import is, which climatology is more suitable for future use in the COLA GCM? To answer this question we examine the seasonal mean systematic errors of surface temperature and precipitation in ensembles CL1 and CL2 in the nine study regions given in Table 6 and shown in Fig. 5b. The 5-yr regional mean observed surface air temperature (T Obs, °C) and observed precipitation (P Obs, mm day⁻¹) in each of these regions is given in Table 7, columns 2 and 5, respectively. The observed surface temperature and precipitation data were obtained from the NCEP Climate Prediction Center Climate Anomaly Monitoring System station data archive (Ropelewski et al. 1985). The corresponding CL1 systematic errors are given in column 3 (T1 Err, °C) and column 6 (P1 Err, mm day⁻¹). The CL2 systematic errors are given in column 4 (T2 Err, °C) and column 7 (P2 Err, mm day⁻¹). In six of the nine regions (regions 3, 7, 8, 9, A, and C) the surface temperature systematic error was reduced in CL2 relative to that in CL1. However, in seven of the nine regions (regions 3, 8, 9, A, B, C, and D) the precipitation systematic error was increased in CL2 relative to that in CL1. Thus, based on the regional systematic error alone, it is difficult to determine which soil wetness climatology results in better seasonal simulations. An examination of the two climatologies for 1 June reveals that the one used in CL1 (Fig. 1a) is rather dry at northern latitudes compared to that used in CL2 (Fig. 5a), as reflected in the mean initial soil wetness differences in regions 3, 8, and 9 (Table 5). These three regions had three of the four worst surface temperature systematic errors in CL1 (2.3°, 3.3°, and 2.3°C), which were all reduced in CL2 (to 0.7°, 1.8°, and 1.5°C). Due to this reduction in what appears to be the worst near-surface boreal summer

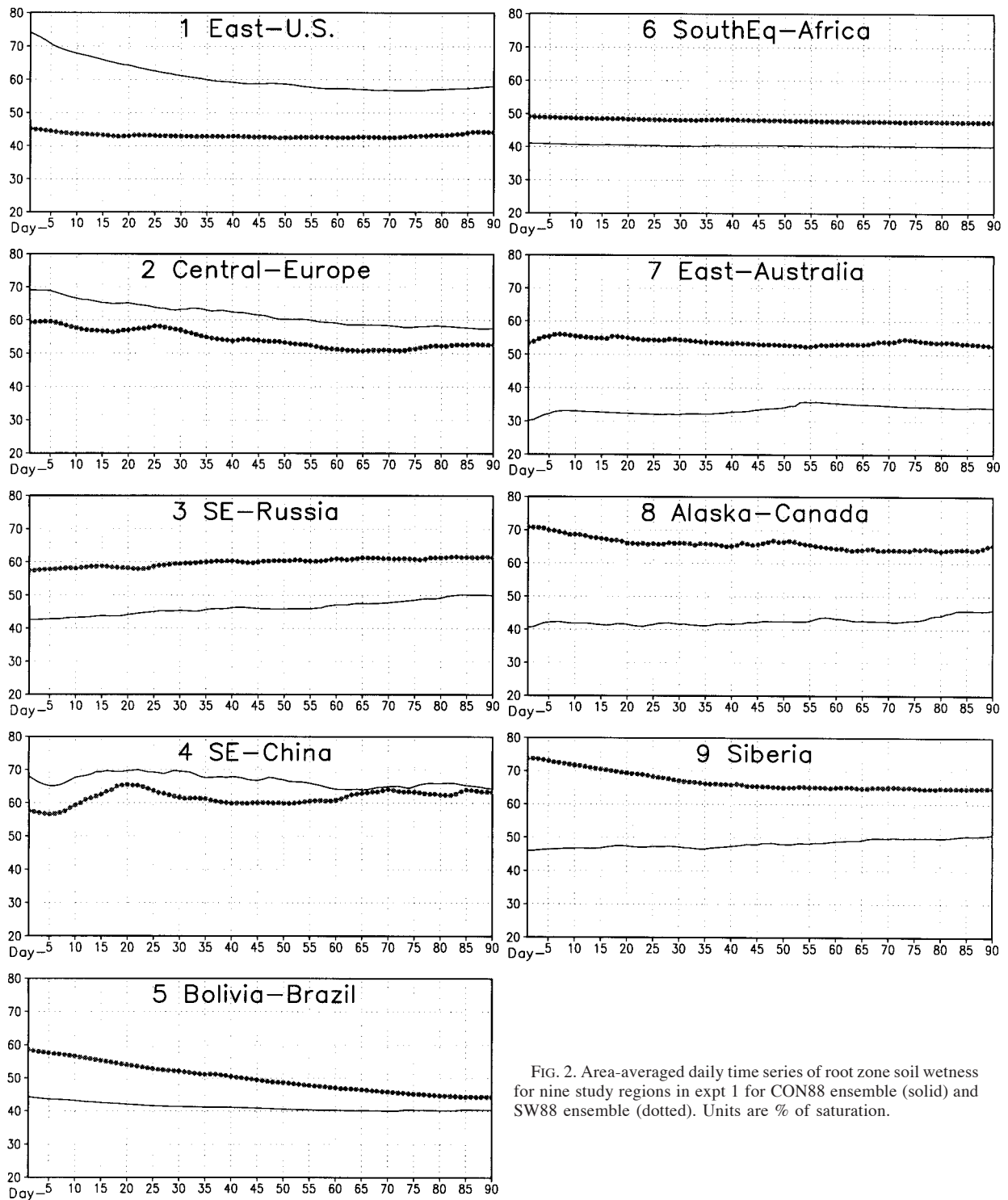


FIG. 2. Area-averaged daily time series of root zone soil wetness for nine study regions in expt 1 for CON88 ensemble (solid) and SW88 ensemble (dotted). Units are % of saturation.

model systematic error, combined with the unusually dry CL1 1 June soil wetness at northern latitudes, the soil wetness climatology used to initialize CL2 was adopted for future use in initializing the COLA GCM.

5. Summary

The intent of this study was to investigate the importance of initial soil wetness in seasonal predictions

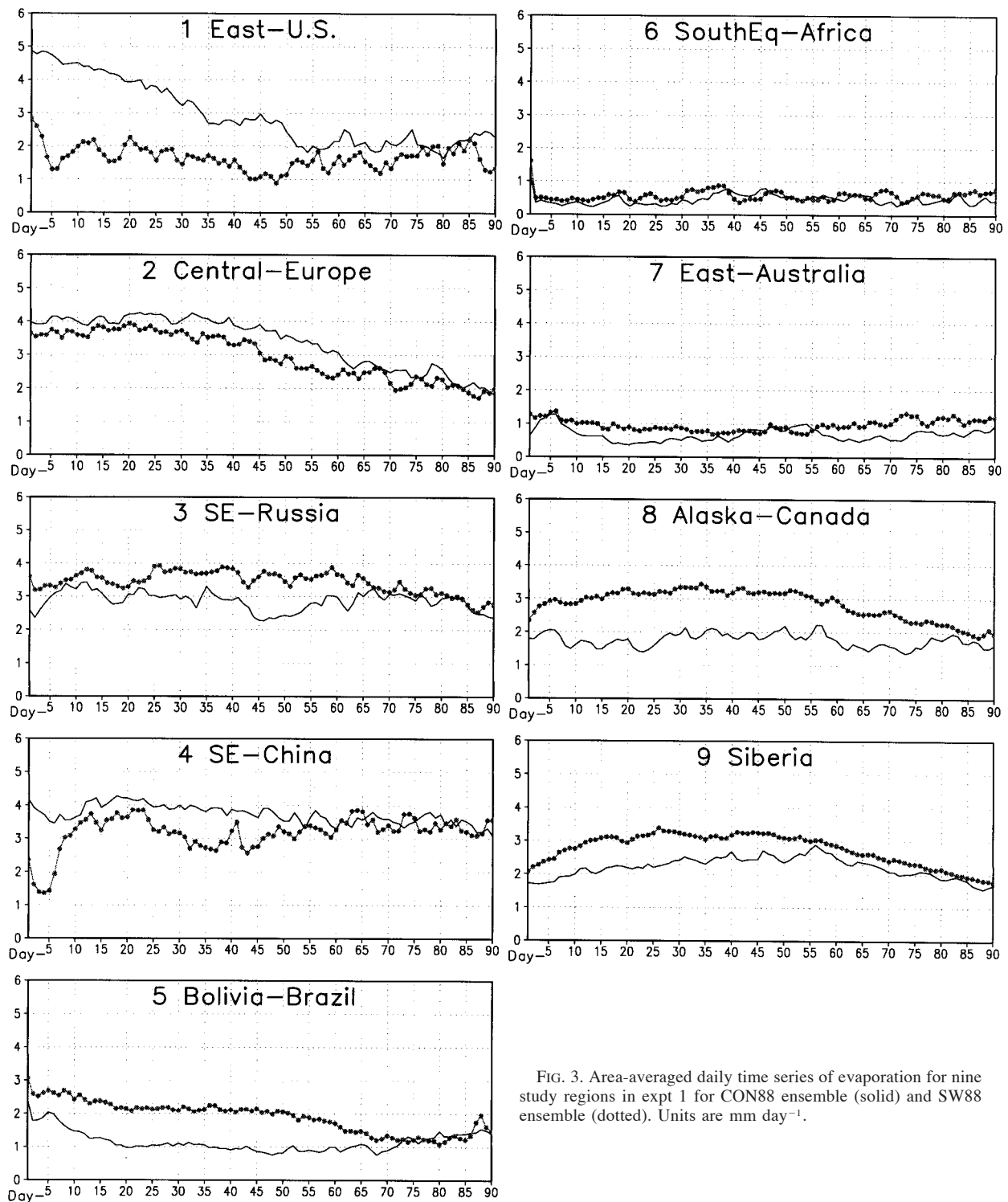


FIG. 3. Area-averaged daily time series of evaporation for nine study regions in expt 1 for CON88 ensemble (solid) and SW88 ensemble (dotted). Units are mm day^{-1} .

with dynamical models. Two experiments were performed, each consisting of two ensembles of global climate model integrations initialized from early June observed atmospheric states. Although the design of the two experiments was somewhat different, both ad-

ressed the basic question of the importance of moderate (and hopefully realistic) soil wetness variations in seasonal prediction. The first experiment examined the impact of using proxy 1988 soil wetness as an initial condition on an ensemble of three 1988 boreal summer

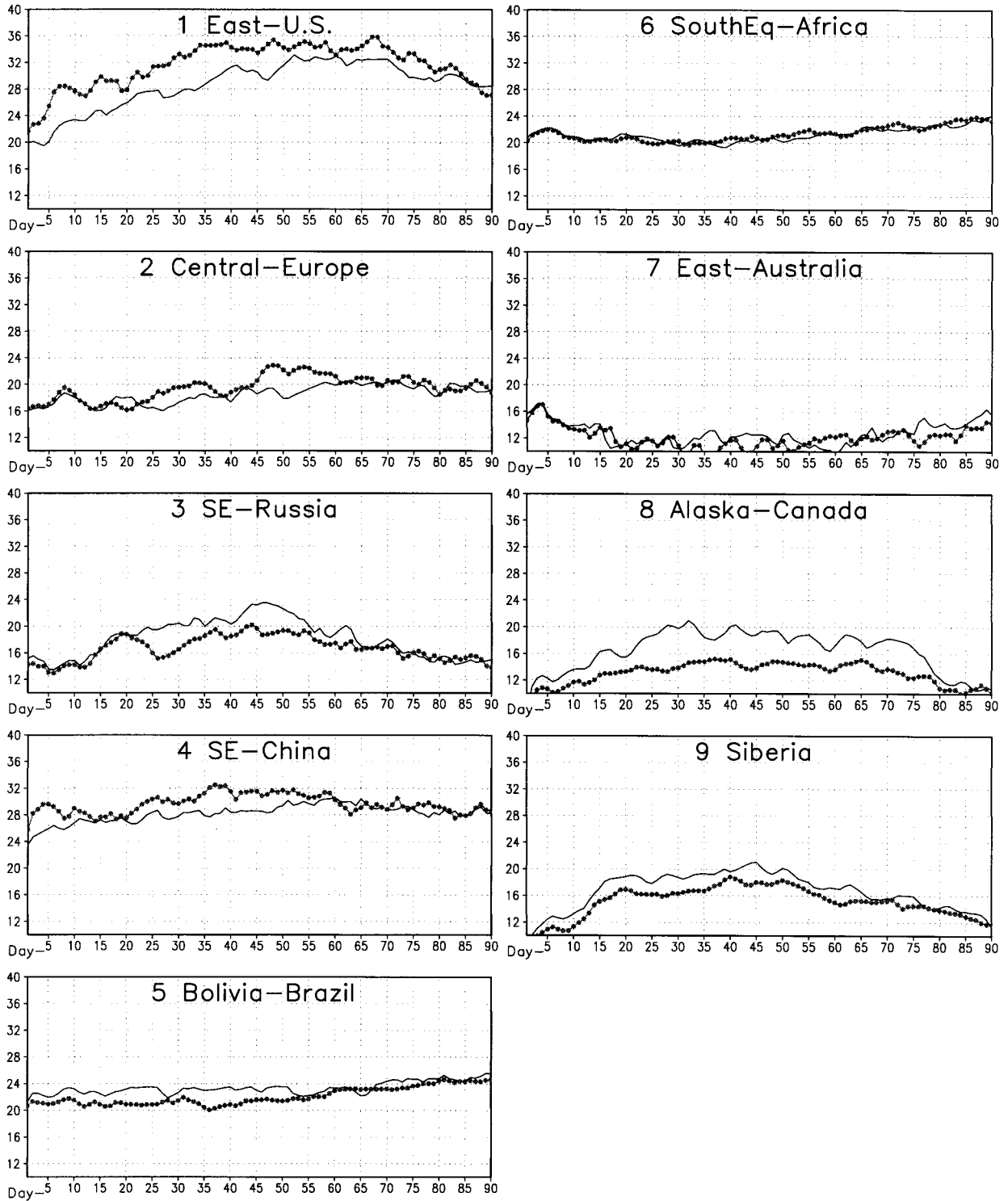


FIG. 4. Area-averaged daily time series of surface temperature for nine study regions in expt 1 for CON88 ensemble (solid) and SW88 ensemble (dotted). Units are °C.

TABLE 4. Integration ensemble names, initiation dates, lengths, SST used and soil wetness initial condition (IC) used in different climatological initial soil wetness experiment.

Ensemble name	Initiation dates	Length (days)	SST	Soil wetness IC
CL1	1 Jun 1987, 1988, 1990, 1992, 1993	90	Observed ^a	Climatological ^b
CL2	1 Jun 1987, 1988, 1990, 1992, 1993	90	Observed ^a	Climatological ^c

^a Reynolds and Smith 1994.

^b From Willmott et al. 1985.

^c 1987–93 climatology from ECMWF analysis–forecast cycle soil moisture.

season integrations. The second experiment examined the impact of using a different soil wetness climatology as an initial condition on an ensemble of boreal summer season integrations from five different years. After initialization, a coupled atmosphere–biosphere model determined the evolution of the soil wetness fields in all

the integrations. The large differences between even the climatological soil wetness fields reflects the considerable uncertainty as to what the observed soil wetness climatology actually is.

In agreement with previously published results, the initial soil wetness impact is found to be largely local and is greatest on near-surface fields. No significant response to the initial soil wetness differences on upper levels of the atmosphere was found. Significant impacts were found in both tropical and extratropical regions in each experiment. The strength of the impact of the initial soil wetness differences, as well as the nature of the impact on precipitation and other atmospheric fields, depends on several factors. These factors include the areal extent and magnitude of the initial soil wetness difference, the persistence of the soil wetness difference, the strength of the solar forcing, the availability of near-by moisture sources, and the strength of the regional dynamical circulation.

Most of the regions with significant increases (decreases) in initial soil wetness had significant increases

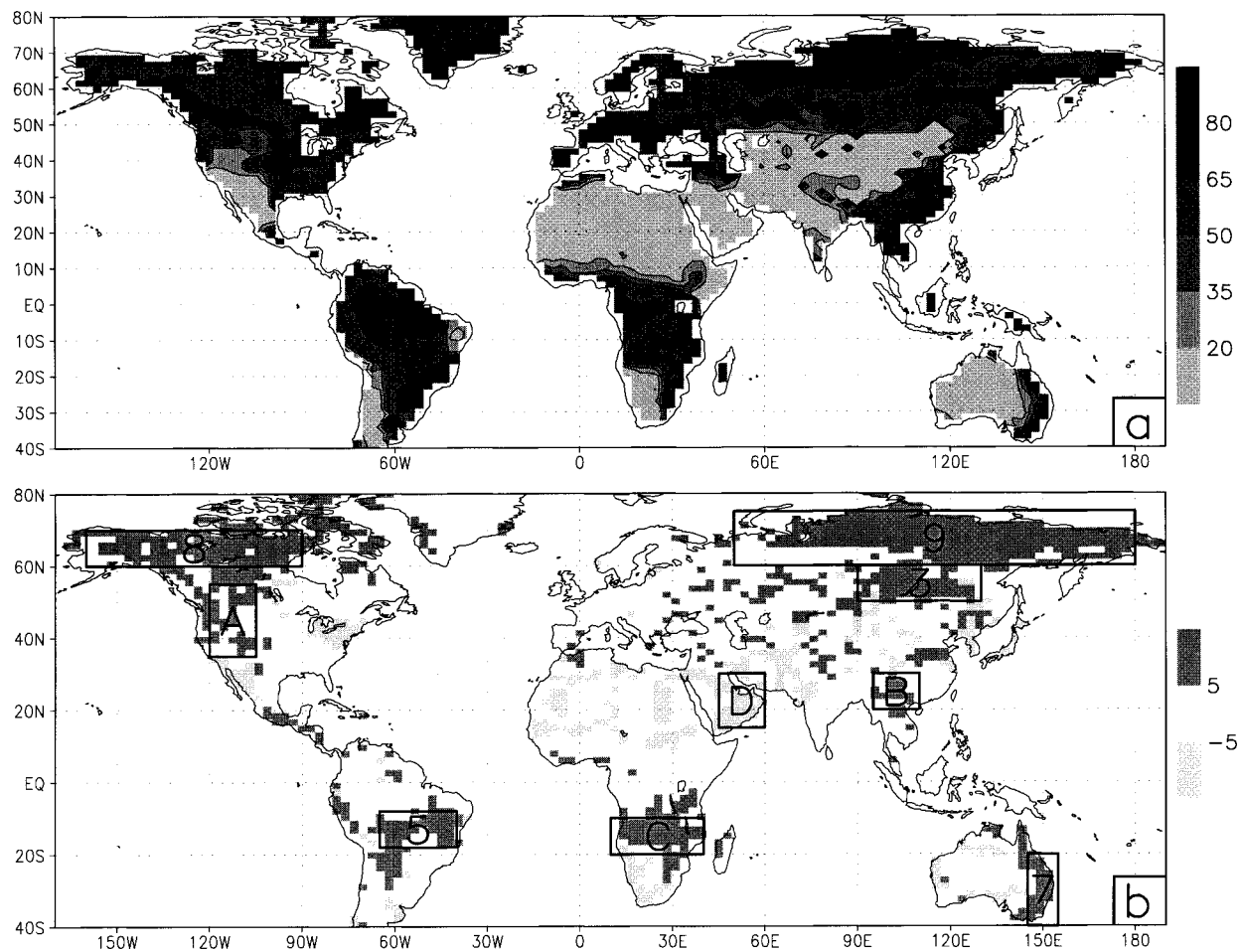


FIG. 5. (a) Soil wetness climatology (% of saturation) for 1 Jun used to initialize CL2 ensemble and (b) initial soil wetness difference in CL2 – CL1 ensembles. Regions with differences greater/less than 10% of saturation are shaded dark/light.

TABLE 5. Study regions, extents and area-averaged initial soil wetness differences (% of saturation) for different climatological initial soil wetness experiment.

#	Region	Bounds	ISW anomalies (%)
3	Southeast Russia	50°–60°N, 90°–130°E	18.8
5	Bolivia–Brazil	8°–18°S, 40°–65°W	15.3
7	East Australia	20°–40°S, 145°–155°E	30.6
8	Alaska–Canada	60°–70°N, 90°–160°W	21.3
9	Siberia	60°–75°N, 50°E–180°	21.1
A	Western United States	35°–55°N, 105°–120°W	13.3
B	Southeast Asia	20°–30°N, 95°–110°E	11.8
C	Southern Africa	10°–20°S, 10°–40°E	16.9
D	Saudi Arabia	15°–30°N, 45°–60°E	–1.4

(decreases) in seasonal mean evaporation and surface air temperature. Considering both experiments as a whole, 16 of 18 study regions had significant evaporation differences and 13 had significant surface temperature differences. The two regions without significant evaporation differences in the second experiment had either a very small initial soil wetness difference (Saudi Arabia, –1.4%) or relatively weak solar forcing (Australia, the southernmost region). Of the five regions without significant surface temperature differences, four had evaporation differences that were less than 0.25 mm day^{–1} in magnitude. The other was eastern Australia in the second experiment, which had a 0.4-mm day^{–1} increase in evaporation that was not significant and that was offset by an equal decrease in the vertically integrated moisture flux convergence. This region was the most “winterlike” of the study regions, with both relatively weak solar forcing and relatively strong local dynamics working against getting a significant response to local soil wetness forcing.

The precipitation response was more variable than that of surface temperature and was more highly dependent on the moisture flux convergence response to the initial soil wetness anomaly. Half of the regions (9 of 18) had significant increases (decreases) in seasonal mean precipitation in response to increased (decreased) initial soil wetness. These nine regions all had seasonal mean evaporation differences that were at least 0.4 mm day^{–1} in magnitude. Only two of the nine regions that had seasonal mean evaporation differences of 0.5 mm day^{–1} magnitude (or greater) did not have a significant precipitation difference of the same sign. In those regions a significant vertically integrated moisture flux difference of the opposite sign completely negated the evaporation difference.

In order to understand why the precipitation was or was not significantly impacted by the initial soil wetness difference, we examined a large number of diagnostics including moisture flux vectors at various levels and vertical profiles of temperature, moisture, and moisture fluxes (not shown). These diagnostics suggest that in order to get a significant increase (decrease) in seasonal

TABLE 6. Area-averaged initial soil wetness difference (ISW, % of saturation) and 90-day mean SW88 – CON88 difference in root zone soil wetness (SWR, % of saturation), evaporation (EVAP, mm day^{–1}), surface temperature (TS, °C), precipitation (P, mm day^{–1}), and vertically integrated moisture flux convergence (VMFC, mm day^{–1}) for the nine study regions in the different climatological initial soil wetness experiment. Bold type indicates area-averaged differences that were significant at the 95% level, as determined by a *t* test.

#	ISW	SWR	EVAP	TS	P	VMFC
3	18.8	13.6	0.6	– 1.5	0.5	–0.1
5	15.3	9.8	0.9	– 1.0	0.0	– 1.0
7	30.6	13.0	0.4	–0.4	–0.1	– 0.4
8	21.3	12.8	0.8	– 1.5	0.8	0.0
9	21.1	11.4	0.4	– 0.8	0.5	0.1
A	13.3	10.0	0.6	– 0.9	0.4	– 0.2
B	11.8	5.1	0.2	–0.2	0.4	0.3
C	16.9	12.7	0.5	– 0.5	0.2	– 0.2
D	–1.4	–0.4	0.1	–0.2	0.1	0.0

mean precipitation, the net of the anomalous vertical and horizontal fluxes of heat and moisture must result in lower levels of the simulated atmosphere (at or below 700 hPa) that are cooler (warmer) and moister (drier). As the upper-atmospheric levels simulated had lesser or negligible differences, the resulting vertical profiles appear more (less) convectively unstable than those obtained without the initial soil wetness difference. Although the notion of convective instability is normally utilized in short-term weather forecasting from analysis of temperature and moisture profiles at a single station, Hao and Bosart (1987) showed that monthly mean area-averaged vertical profiles of temperature and dewpoint calculated from observations reflected the anomalously low convective instability during a dry summer. We examined the ensemble seasonal mean vertical profiles of temperature and dewpoint (and their differences) for each of the 18 regions in the two experiments. Only in the nine cases that had significant precipitation differences was a coherent change in the dewpoint depression (temperature – dewpoint temperature) evident. To summarize this change we calculated the mean of the difference in dewpoint depression at 700, 850, and 1000 hPa due to the change in initial soil wetness for each

TABLE 7. Areal averaged 5-yr mean observed surface temperature (T Obs, °C) and precipitation (P Obs, mm day^{–1}), CL1 systematic errors in surface temperature (T1 Err, °C) and precipitation (P1 Err, mm day^{–1}), and CL2 systematic errors in surface temperature (T2 Err, °C) and precipitation (P2 Err, mm day^{–1}).

#	T Obs	T1 Err	T2 Err	P Obs	P1 Err	P2 Err
3	14.3	2.3	0.7	2.6	1.5	2.0
5	24.2	–0.9	–1.8	0.7	0.1	0.0
7	12.7	2.3	2.0	1.8	1.1	1.0
8	9.4	3.3	1.8	1.3	1.3	2.0
9	9.3	2.3	1.5	1.3	1.1	1.6
A	17.8	0.8	–0.1	1.4	1.4	1.7
B	25.0	–3.4	–3.6	6.9	1.4	1.9
C	19.0	1.1	0.6	0.3	0.1	0.4
D	33.1	–0.5	–0.7	0.1	0.4	0.6

region. This mean dewpoint depression difference was highly negatively correlated with the precipitation difference, with a correlation coefficient of -0.88 over the 18 regions. Moreover, the nine regions with significant positive precipitation differences all had dewpoint depression differences one to two degrees in magnitude; the remaining nine regions had dewpoint depression differences half a degree or less in magnitude. Thus, in order for an initial soil wetness difference to force a significant seasonal mean precipitation difference in a region, it must effectively alter the mean convective stability of the region. This result is consistent with the simulated regional precipitation differences, which all occurred in the convective component of the precipitation.

The proxy 1988 soil wetness initial state resulted in significant seasonal mean evaporation, surface temperature, precipitation, and sea level pressure differences in the eastern United States in the 1988 ensembles, which are indicative of drought conditions similar to those observed during the 1988 U.S. drought. This implies that dry soil conditions set up by the late spring-early summer drought (precipitation deficit) may have contributed to the dry and hot summer conditions observed over the United States during 1988. We should note that this was a highly anomalous year and that the influence of initial soil wetness for other years may be quite different.

An examination of the model surface air temperature systematic errors revealed that the worst model boreal summer systematic biases in the ensemble initialized with climatological soil wetness by Willmott et al. (1985) were in northern latitudes where the soil appeared to be unrealistically dry for 1 June. These biases were greatly reduced in the ensemble initialized with climatological soil wetness derived from the ECMWF analysis-forecast system prognostic soil moisture for 1987-93. On this basis the later climatology was chosen for future integrations.

Although the results presented here may be somewhat model dependent, they strongly suggest that seasonal atmospheric prediction could be enhanced by using a realistic initial state of soil wetness.

Acknowledgments. The authors would like to thank T. Palmer and Č. Branković for their help in obtaining and interpreting the ECMWF soil wetness analysis, P. Sellers for his guidance in adapting the ECMWF soil wetness, and R. Reynolds for providing the SST analyses. This research was supported by NASA Grants NAGW-1269 and NAGW-2661 and by NSF Grant ATM-90-19296.

REFERENCES

- Alpert, J. C., M. Kanamitsu, P. M. Caplan, J. G. Sela, G. H. White, and E. Kalnay, 1988: Mountain induced gravity wave drag parameterization in the NMC medium-range forecast model. Preprints, *Eighth Conf. on Numerical Weather Prediction*, Baltimore, MD, Amer. Meteor. Soc., 726-733.
- Anthes, R. A., 1977: A cumulus parameterization scheme utilizing a one-dimensional cloud model. *Mon. Wea. Rev.*, **105**, 270-300.
- Arakawa, A., and W. H. Schubert, 1974: Interaction of cumulus cloud ensemble with the large-scale environment. *J. Atmos. Sci.*, **31**, 671-701.
- Atlas, R., N. Wolfson, and J. Terry, 1993: The effect of SST and soil moisture anomalies on GLA model simulations of the 1988 U.S. summer drought. *J. Climate*, **6**, 2034-2048.
- Betts, A. K., J. H. Ball, A. C. M. Beljaars, M. J. Miller, and P. Viterbo, 1994: Coupling between land surface boundary layer parameterizations and rainfall on local and regional scales: Lessons from the wet summer of 1993. Preprints, *Fifth Symp. on Global Change*, Nashville, TN, Amer. Meteor. Soc., 174-181.
- Branković, Č., T. N. Palmer, and P. Viterbo, 1990: Diagnostic and model sensitivity study of the 1988 U.S. drought. *Proc. Workshop on 1988 U.S. Drought*, College Park, MD, Dept. of Meteorology, University of Maryland at College Park, 110-128. [Available from Dept. of Meteorology, University of Maryland at College Park, College Park, MD 20742.]
- Cook, K. H., and A. Gnanadesikan, 1991: Effects of saturated and dry land surfaces on the tropical circulation and precipitation in a general circulation model. *J. Climate*, **4**, 873-889.
- Delworth, T. L., and S. Manabe, 1988: The influence of potential evaporation on the variabilities of simulated soil wetness and climate. *J. Climate*, **1**, 523-547.
- , and —, 1989: The influence of soil wetness on near-surface atmospheric variability. *J. Climate*, **2**, 1447-1462.
- Dewitt, D., 1996: The effects of cumulus convection parameterization on the climate of the COLA general circulation model. COLA Tech. Rep. 27, 43 pp. [Available from COLA, 4041 Powder Mill Road, Suite 302, Calverton, MD 20705.]
- Dirmeyer, P. A., 1994: Vegetation stress as a feedback mechanism in midlatitude drought. *J. Climate*, **7**, 1463-1483.
- , and J. Shukla, 1993: Observational and modeling studies of the influence of soil moisture anomalies on the atmospheric circulation. *Predictions of Interannual Climate Variations*, J. Shukla, Ed., NATO Series I, Vol. 6, Springer-Verlag, 1-23.
- Dorman, J. L., and P. J. Sellers, 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**, 833-855.
- Fennessy, M. J., J. L. Kinter III, L. Marx, E. Schneider, P. J. Sellers, and J. Shukla, 1994: GCM simulations of the life cycles of the 1988 US drought and heat wave. COLA Tech. Rep. 6, 53 pp. [Available from COLA, 4041 Powder Mill Road, Suite 302, Calverton, MD 20705.]
- Hao, W., and L. F. Bosart, 1987: A moisture budget analysis of the protracted heat wave in the southern plains during the summer of 1980. *Wea. Forecasting*, **2**, 269-288.
- Harshvardhan, R. Davies, D. A. Randall, and T. G. Corsetti, 1987: A fast radiation parameterization for general circulation models. *J. Geophys. Res.*, **92**, 1009-1016.
- Hogan, T. F., and T. E. Rosmond, 1991: The description of the Navy Operational Global Atmospheric Prediction System's spectral forecast model. *Mon. Wea. Rev.*, **119**, 1786-1815.
- Hou, Y.-T., 1990: Cloud-radiation-dynamics interaction. Ph.D. thesis, University of Maryland at College Park, 209 pp. [Available from Dept. of Meteorology, University of Maryland at College Park, College Park, MD 20742.]
- Kinter, J. L., III, J. Shukla, L. Marx, and E. K. Schneider, 1988: A simulation of the winter and summer circulations with the NMC global spectral model. *J. Atmos. Sci.*, **45**, 2486-2522.
- Kuo, H. L., 1965: On the formation and intensification of tropical cyclones through latent heat release by cumulus convection. *J. Atmos. Sci.*, **22**, 40-63.
- Lacis, A. A., and J. E. Hansen, 1974: A parameterization for the absorption of solar radiation in the earth's atmosphere. *J. Atmos. Sci.*, **31**, 118-133.

- Liston, G. E., Y. C. Sud, and G. K. Walker, 1993: Design of a global soil moisture initialization procedure for the Simple Biosphere Model. NASA Tech. Memo. 104590, 130 pp. [Available from NTIS, 5285 Port Royal Rd., Springfield, VA 22161; NTIS N94-1094012INZ.]
- Mellor, G. L., and T. Yamada, 1982: Development of a turbulence closure model for geophysical fluid problems. *Rev. Geophys. Space Phys.*, **20**, 851–875.
- Mintz, Y., 1984: The sensitivity of numerically simulated climates to land-surface conditions. *The Global Climate*, J. Houghton, Ed., Cambridge University Press, 79–105.
- , and Y. Serafini, 1981: Global fields of soil moisture and land-surface evapotranspiration. Research review—1980/81. NASA/Goddard Space Flight Center Tech. Memo. 83907, 269 pp. [Available from NASA/Goddard Space Flight Center, Greenbelt, MD 20771.]
- , and —, 1992: A global monthly climatology of soil moisture and water balance. *Climate Dyn.*, **8**, 13–17.
- Miyakoda, K., and J. Sirutis, 1977: Comparative integrations of global spectral models with various parameterized processes of subgrid scale vertical transports. *Beitr. Phys. Atmos.*, **50**, 445–447.
- , —, and R. F. Strickler, 1979: Cumulative results of extended forecast experiments. Part II: Model performance for summer cases. *Mon. Wea. Rev.*, **107**, 395–420.
- Moorthi, S., and M. J. Suarez, 1992: Relaxed Arakawa–Schubert: A parameterization of moist convection for general circulation models. *Mon. Wea. Rev.*, **120**, 978–1002.
- Oglesby, R. J., 1991: Springtime soil moisture, natural climatic variability, and North American drought as simulated by the NCAR Community Climate Model 1. *J. Climate*, **4**, 890–897.
- , and D. J. Erickson III, 1989: Soil moisture and the persistence of North American drought. *J. Climate*, **2**, 1362–1380.
- Paegle, J., K. C. Mo, and J. Nogués-Paegle, 1996: Dependence of simulated precipitation on surface evaporation during the 1993 United States summer floods. *Mon. Wea. Rev.*, **124**, 345–361.
- Phillips, N. A., 1957: A coordinate system having some special advantages for numerical forecasting. *J. Meteor.*, **14**, 184–185.
- Reynolds, R. W., 1988: A real-time global sea surface temperature analysis. *J. Climate*, **1**, 75–86.
- , and T. M. Smith, 1994: Improved global sea surface temperature analyses using optimum interpolation. *J. Climate*, **7**, 929–948.
- 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.
- Ropelewski, C. F., J. E. Janowiak, and M. F. Halpert, 1985: The analysis and display of real time surface climate data. *Mon. Wea. Rev.*, **113**, 1101–1107.
- Sato, N., P. J. Sellers, D. A. Randall, E. K. Schneider, J. Shukla, J. L. Kinter III, Y.-T. Hou, and E. Albertazzi, 1989a: Effects of implementing the Simple Biosphere Model in a general circulation model. *J. Atmos. Sci.*, **46**, 2757–2782.
- , —, —, —, —, —, —, and —, 1989b: Implementing the Simple Biosphere Model (SiB) in a GCM: Methodology and results. NASA Contractor Report CR-185509, NASA, Washington, DC, 76 pp. [Available from NTIS, 5285 Port Royal Rd., Springfield, VA 22161.]
- Sela, J. G., 1980: Spectral modeling at the National Meteorological Center. *Mon. Wea. Rev.*, **108**, 1279–1292.
- Sellers, P. J., 1990: Specifying surface boundary conditions for drought studies. *Proc. Workshop on 1988 U.S. Drought*, College Park, MD, Dept. of Meteorology, University of Maryland at College Park, 170–172. [Available from Dept. of Meteorology, University of Maryland at College Park, College Park, MD 20742.]
- , Y. Mintz, Y. C. Sud, and A. Dalcher, 1986: A Simple Biosphere Model (SiB) for use within general circulation models. *J. Atmos. Sci.*, **43**, 505–531.
- , W. J. Shuttleworth, J. L. Dorman, A. Dalcher, and J. M. Roberts, 1989: Calibrating the simple biosphere model for Amazonian tropical forest using field and remote sensing data. Part I: Average calibration with field data. *J. Appl. Meteor.*, **28**, 727–759.
- Shukla, J., and Y. Mintz, 1982: Influence of land-surface evapotranspiration on the earth's climate. *Science*, **215**, 1498–1501.
- Slingo, J. M., 1987: The development and verification of a cloud prediction scheme for the ECMWF model. *Quart. J. Roy. Meteor. Soc.*, **113**, 899–927.
- Sud, Y. C., and M. J. Fennessy, 1984: Influence of evaporation in semi-arid regions on the July circulation: A numerical study. *J. Climatol.*, **4**, 383–398.
- Tiedtke, M., 1984: The effect of penetrative cumulus convection on the large-scale flow in a general circulation model. *Beitr. Phys. Atmos.*, **57**, 216–239.
- Vernekar, A., B. Kirtman, J. Zhou, and D. Dewitt, 1992: Orographic gravity-wave drag effects on medium-range forecasts with a general circulation model. *Physical Processes in Atmospheric Models*, D. R. Sikka and S. S. Singh, Eds., Wiley Eastern Limited, 295–307.
- Walker, J., and P. R. Rowntree, 1977: The effect of soil moisture on circulation and rainfall in a tropical model. *Quart. J. Roy. Meteor. Soc.*, **103**, 29–46.
- Willmott, C. J., C. M. Rowe, and Y. Mintz, 1985: Climatology of the terrestrial seasonal water cycle. *Int. J. Climatol.*, **5**, 589–606.
- Xue, Y., P. J. Sellers, J. L. Kinter, and J. Shukla, 1991: A simplified biosphere model for global climate studies. *J. Climate*, **4**, 345–364.