

The Predictability of Winter Snow Cover over the Western United States

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ABSTRACT

A set of model runs was made with the National Center for Atmospheric Research (NCAR) Community Climate Model, version 3 (CCM3) to investigate and help assess the relative roles of snow cover anomalies and initial atmospheric states on the subsequent accumulation and ablation seasons. In order to elucidate the physical mechanisms responsible for the large impact in one case but small impact in the other, two experiments with CCM3 were made that imposed an exaggerated initial snow cover [1-m snow water equivalent (SWE)] over the western U.S. domain. One run was started on 1 December, the other on 1 February. These runs made it clear that the high albedo of snow was the dominant physical process. An additional set of runs with realistic yearly snow anomalies was also made. Results suggest that for runs starting in February (late winter), the initial prescription of snow cover is more important than the initial atmospheric state in determining the subsequent evolution of snow cover. For runs starting in December (early winter), the results are less clear, with neither the initial snow cover nor the initial state of the atmosphere appearing to be the dominant factor. In February, when the sun is relatively high in the sky and days are longer, the albedo effect is a dominant factor; while in December the sun was too low in the sky and days too short for the albedo effect to be important. As the winter season progressed, the subsequent accumulation of snow eliminated the effects of the initial December anomalies.

1. Introduction

The effect of atmosphere–snow interactions on the climate of midlatitude continents, including the western United States, has long been a topic of speculation and study, but remains poorly understood. Many previous studies have shown that snow cover affects lower-tropospheric temperatures (Wagner 1973; Dewey 1977; Baker et al. 1992; Walsh et al. 1982; Leathers et al. 1995) and atmospheric circulation (Dickson and Namais 1976; Heim and Dewey 1984; Clark and Serreze 2000) on regional and possibly even hemispherical scales. The depression of near-surface air temperatures found over snow cover is generally attributed to the higher albedo of the land surface when snow cover is present. This albedo–temperature feedback is largest in the spring when snow cover remains extensive and insolation is high (Groisman et al. 1994). Snow cover also lowers

tropospheric air temperatures by redirecting surface energy inputs toward warming and melting/sublimating the snowpack (Groisman et al. 1997). This acts as a stabilizing mechanism for the atmosphere, and in conjunction with radiative cooling from the snow-covered surface can induce strong temperature inversions in the boundary layer. Snow cover over Eurasia has been shown to affect the strength and onset of the Asian summer monsoon and may appear to influence the strength of the monsoonal circulation over the western United States (see e.g., Foster et al. 1983; Barnett et al. 1989; Gutzler and Preston 1997; Cayan 1996). Accurate analysis and ultimately, prediction of atmosphere–snow cover interactions on seasonal to longer timescales, requires detailed understanding of the competing roles of local effects (the snow cover), remote forcings [e.g., tropical sea surface temperature (SST) anomalies], and internal variability.

Since previous work makes it clear that snow cover can act as a climatic forcing mechanism, some degree of seasonal climate predictability may be inherent in snow cover. That is, knowing the initial state of the snow

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cover may enable some prediction skill in forecasting the subsequent evolution of snow cover and hence its impact on the atmosphere. The most basic question to answer is whether the initial state of the snow cover is positively or negatively correlated with the subsequent snow cover. In other words, does a large initial snow cover (in extent and volume) tend to perpetuate itself, or rather to instead induce feedbacks that limit its eventual size? Or is there no strong correlation at all, which implies that atmospheric variability is more important than the initial state of the atmosphere? We focus on these questions, and therefore have constructed ensembles of predictability studies using the National Center for Atmospheric Research (NCAR) Community Climate Model, version 3 (CCM3) in which we compared the relative roles of initial surface and atmospheric conditions over the western U.S. region in determining the subsequent evolution of snow cover. We have also made sensitivity studies with exaggerated snow cover anomalies in order to determine the physical processes and linkages with the atmosphere that may be important in accounting for this predictability.

2. Model description—NCAR CCM3

The global climate model used for this study is the NCAR Climate System Model (CSM; Boville and Gent 1998). The CSM includes atmospheric, oceanic, land surface, and sea ice components. The atmospheric component of the model (CCM3) has a horizontal resolution of T42, or an equivalent grid resolution of 2.8° latitude by 2.8° longitude, and the vertical is resolved by 18 layers. The standard land surface option for CSM (and hence CCM3) is the Land Surface Model (LSM; Bonan 1998). Soil and vegetation type and characteristics are prescribed and vary monthly. Soil temperatures and soil moisture are calculated using a six-layer soil energy and moisture model. The LSM incorporates components of the snow hydrology of Marshall and Oglesby (1994), including a variable snow cover albedo and a surface albedo based on fractional snow/vegetation cover. Snow water equivalent is determined using a simple mass and energy balance and assuming a constant snow density. Kiehl et al. (1998) and Hack et al. (1998) summarize important characteristics of the model-generated climate.

A baseline model simulation used in this study is a 42-yr CCM3 run with monthly SST for each year specified according to observations supplied by the National Centers for Environmental Prediction (NCEP) for 1958–99 (henceforth called CCM3/SST). Results from the CCM3/SST run can be directly compared to atmospheric observations on a year-to-year basis, with the caveat that the SST forcing is the only forcing that relates model years to actual calendar years (Oglesby et al. 2001). This run is used to provide the interannual snow cover anomalies used for the present study, as well as to serve as a “control” against which to gauge the degree of any predictability.

3. Observed and modeled snow cover over the western United States

Brown (2000) describes snow cover variability and change over the Northern Hemisphere during most of the twentieth century. He found a large degree of variability over North America as a whole, with a positive long-term trend. Most of the stations used in this study were from Canada and the eastern half of the United States so it is not clear how well these conclusions apply to the western United States. Serreze et al. (1999) used daily station data from the U.S. Department of Agriculture snowpack telemetry sites throughout the western United States to characterize regional patterns of snow accumulation and ablation. These daily records, mostly from the period 1979–96, were used to develop regional climatologies of snow water equivalent (SWE) and ratios of SWE to rainfall. While the period of record was too short to identify any trends, their study of the variability of SWE and total precipitation showed a high degree of sensitivity to temperature, especially along the Pacific coast and in Arizona–New Mexico. In related work, Serreze et al. (2001) used snowpack telemetry (SNOTEL) data to examine the frequency and context of large snowfall events. They found that large events in the marine regions were most common in midwinter, whereas for the Rocky Mountain regions, large events were more common in the spring. Furthermore, they found a significant positive correlation between seasonal snowfall anomalies and the number of large events and a relationship between the number of large events and El Niño–La Niña conditions.

Yang and Niu (2002, manuscript submitted to *J. Climate*) performed an assessment of the ability of the LSM to model the snow cover climatology. They find the LSM snow model replicates reasonably well the melt rates of snow cover in the spring but does more poorly during the accumulation season. This is partly due to a warm bias over snow cover found in the LSM. Although the LSM underestimates peak snow accumulation, it overestimates total snow mass over North America and Eurasia when compared with estimates from the *Nimbus-7* Scanning Multichannel Microwave Radiometer (SMMR).

Such passive microwave measurements of snow water equivalent are inaccurate for mapping SWE in regions where there is even moderate amounts of vegetation present and where snow grain metamorphism is active. Furthermore, when the snow is melting, passive microwave measurements cannot be used at all. Therefore, we compare the frequency of snow cover occurrence between observations and the CCM3 for December and February (Figs. 1 and 2). The observations were obtained from the snow cover portion of the *Northern Hemisphere EASE-Grid Weekly Snow Cover and Sea Ice Extent, Version 2* (Armstrong and Brodzik 2002). The original weekly snow data are the National Oceanic and Atmospheric Administration (NOAA) weekly snow

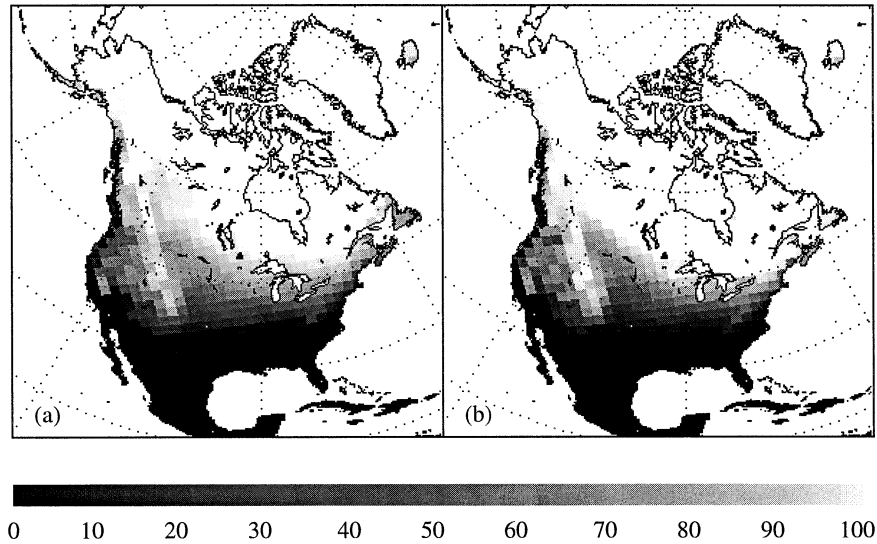


FIG. 1. Climatology of (a) Dec and (b) Feb satellite-derived North American snow cover frequency (%). Data are from the *Northern Hemisphere EASE-Grid Weekly Snow Cover and Sea Ice Extent, Version 2* (1966–2001).

maps, regridded by National Snow and Ice Data Center (NSIDC) into the 25-km Equal-Area Scalable Earth Grid (EASE-Grid) using a nearest neighbor interpolation. The period of record is approximately October 1966 through June 2001. Data values range from 0.0 (no snow at this pixel in this month, for the period of record) to 1.0 (this pixel is always snow-covered during this month for the period of record). Model output was processed daily for the 42 yr of the CCM3/SST run.

Comparison of Figs. 1 and 2 show that the CCM3 and observed snow frequencies agree fairly well over North America for both December and February. Most of Canada has frequency of coverage 80% or greater, as do the Rocky Mountains of the western United States.

The southern and central regions of the United States have very low (<20%) frequency of occurrence. In the model results, this region of very low frequency extends well south of what is seen in the observations (this may be due, in part, to the difference in temporal scales used to determine the snow frequency). Note also the northward bulge of low frequency occurrence from the U.S. Great Plains into the Canadian prairie in both the model and observations. These results suggest that the CCM3 does a reasonably good job at simulating where and how often snow cover should occur.

Assessment of snow depths in the model is more problematical, largely due to lack of observational coverage, and difficulties in transforming geometric snow

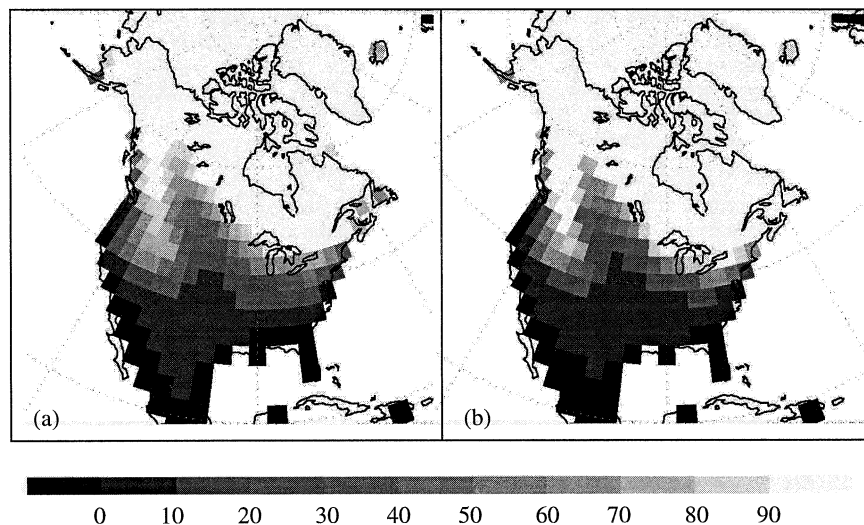


FIG. 2. Climatology of (a) Dec and (b) Feb model-derived North American snow cover frequency (%). Data are from the 42-yr CCM3/SST simulation.

depth into SWE. Most meteorological reporting stations in the western United States are located in valleys, and hence biased toward low snow, while the SNOTEL stations are likely biased toward mountain locations that may not be well resolved at the fairly coarse resolution of the model. We have examined SNOTEL stations over the western U.S. region and found that even after attempting crude elevation corrections, CCM3 tends to have thinner snowpacks and ones that melt off somewhat more quickly than seen in the observations from SNOTEL. This may also explain some of the relatively lower frequencies found in the CCM3 plots over the western United States, although the model does not appear seriously deficient in this regard.

The extent of the western U.S. domain we use is arbitrary, and chosen to be large enough that it might be expected to have an impact on a relatively coarse atmospheric GCM, while still being small enough to focus on a specific region. Obviously, questions of spatial scale arise in considering snow cover–atmosphere interactions, but our goal here is elucidating physical mechanisms, and evaluating the potential for predictability based on snow cover. Spatial and temporal variability across a region as large as the western United States may also be important. For runs (described later) in which we take actual model anomalies as initial conditions, this variability is preserved, at least initially. For the runs with an unrealistically large initial anomaly, as described later, the goal is to determine the maximal extent of snow–atmosphere interactions. We do note that much of the center of our region contains the Rocky Mountains, which have a very high frequency (>70%) of snow cover, with surrounding regions on all sides except the north having a much lower frequency (generally <50%) of snow cover.

4. Experiment design

We have designed two types of experiments to investigate the predictability of midwinter and late spring snow cover and the feedback mechanisms responsible. Each set of experiments is described below in detail. Some studies suggest that SST anomalies, such as those connected to the ENSO events, have an impact on winter precipitation (and hence snow cover) over regions in the western United States (e.g., Smith and O'Brien 2001; Kunkel and Angel 1999; Janowiak and Bell 1998). An analysis of CCM3/SST winter precipitation anomalies and November–May monthly snow cover anomalies of dates of warm and cold phase El Niño–Southern Oscillation (ENSO) events shows no significant connection in regions of the western United States. Therefore, the role of SSTs in predictability of snow cover anomalies was pursued in this paper only to the extent to which they affected the initial atmospheric states of the simulations.

a. ANOM experiments

To clarify the physical mechanisms by which snow cover can affect climate, we made two simulations in which an extreme anomaly of 1-m SWE is imposed over the western United States domain only (see Figs. 3d and 4d). The first simulates the climate response to an initial snow anomaly imposed on 1 February (hereafter, ANOM-FEB), while the second simulates the climate response to an anomaly imposed on 1 December (hereafter, ANOM-DEC).

The February runs represent mid to late winter conditions, that is, in the midst of the snow season, and more importantly, a time when the sun is relatively high in the sky and days are becoming longer as the earth's annual cycle progresses away from the Northern Hemisphere winter solstice. This should result in a relatively large albedo effect. The December runs represent late fall/early winter snow conditions (i.e., near the onset of the snow season) and, importantly, a time near the winter solstice when the sun is low in the sky and period of daylight short. This means the albedo effect of snow cover is minimized. Other processes, namely, the high longwave emissivity of snow cover and consequent strong surface cooling, and the ability of snow cover to insulate the underlying surface, should be largely unaffected by the time of year. Thus, these paired experiments directly assess the effects of high snow reflectivity (albedo).

The objective of these runs is to analyze the physical mechanisms by which any adjustment of the atmosphere occurs given an anomalous snow cover. We choose to impose an exaggerated anomaly to more clearly identify these mechanisms. One meter of initial SWE will persist for a considerable period of time, even during strong snowmelt conditions. These runs also allow us to distinguish the effects of a snow cover anomaly just over the western U.S. domain as opposed to anomalies that also appear elsewhere around the globe.

The anomalies are imposed on a model year that otherwise corresponds to a relatively "low" snow cover anomaly (1969 for the February simulation and 1972 for the December simulation, see Tables 1 and 2 and Fig. 5). Both experiments were begun on the first day of the simulation month (either 1 February or 1 December) and were run through the following July to allow for much of the imposed snow cover to ablate. Even so, anomalous snow cover still persists in the ANOM-FEB experiment well into July.

b. SPRED experiments

We made two sets of snow "predictability experiments" (SPRED), each set containing five simulations (see Tables 1 and 2). One set was based on years that had relatively high or low January, February, and March snow anomalies over the western United States and that clearly demonstrated these anomalies on 1 February,

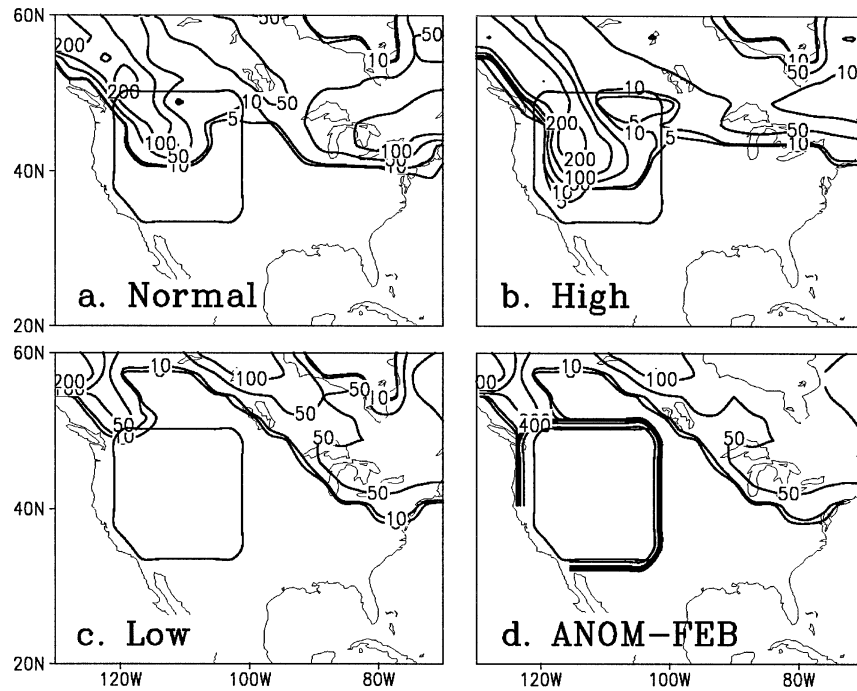


FIG. 3. SWE (mm) for 1 Feb showing (a) the “normal” initial conditions of snow cover (1 Feb 1971 of CCM3/SST simulation), (b) an anomalously high snow cover (1 Feb 1967 of CCM3/SST simulation), (c) an anomalously low snow cover (1 Feb 1969 of CCM3/SST simulation), and (d) the initial snow cover for the 1-m snow experiment (ANOM-FEB). The model domain is indicated by the trapezoidal outline.

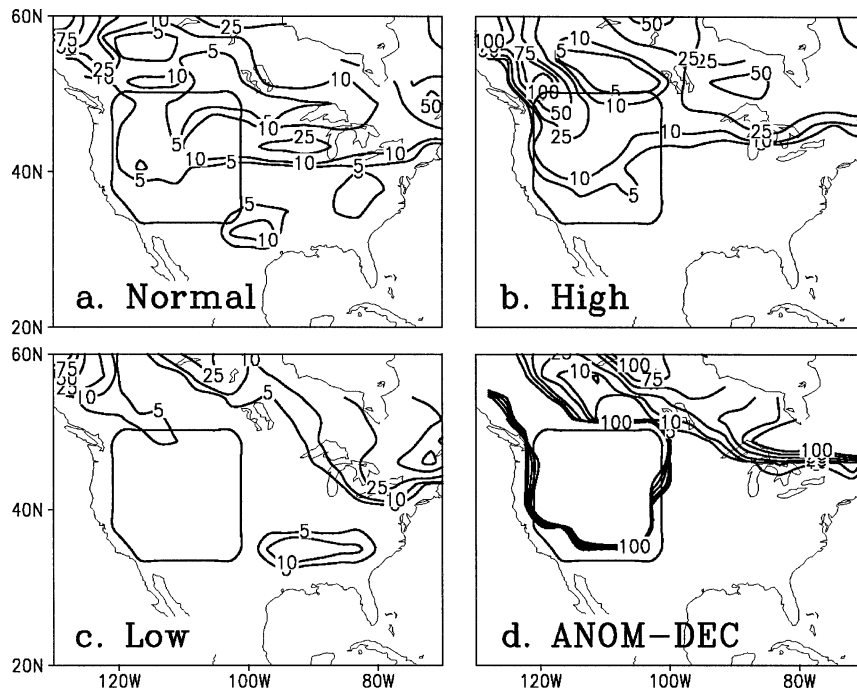


FIG. 4. SWE (mm) for 1 Dec showing (a) the “normal” initial conditions of snow cover (1 Dec 1985 of CCM3/SST simulation), (b) an anomalously high snow cover (1 Dec 1963 of CCM3/SST simulation), (c) an anomalously low snow cover (1 Dec 1972 of CCM3/SST simulation), and (d) the initial snow cover for the 1-m snow experiment (ANOM-DEC). The model domain is indicated by the trapezoidal outline.

TABLE 1. Basic predictability experiments for Feb start date.

Experiment	Initial conditions		Start date	Length (days)
	Atmosphere	Snow		
SPRED1	Normal ^a	High ^b	1 Feb	90
SPRED2	High	Normal	1 Feb	90
SPRED3	Normal	Low ^c	1 Feb	90
SPRED4	Low	Normal	1 Feb	90
ANOM-FEB	Low	1-m SWE	1 Feb	181

^a Normal snow = 1 Feb 1971 initial conditions.

^b High snow = 1 Feb 1967 initial conditions with an avg initial snow cover anomaly of +32.2 mm SWE relative to normal.

^c Low snow = 1 Feb 1969 initial conditions with an avg initial snow cover anomaly of -21.8 mm SWE relative to normal.

TABLE 2. Basic predictability experiments for Dec start date.

Experiment	Initial conditions		Start date	Length (days)
	Atmosphere	Snow		
SPRED5	Normal ^a	High ^b	1 Dec	90
SPRED6	High	Normal	1 Dec	90
SPRED7	Normal	Low ^b	1 Dec	90
SPRED8	Low	Normal	1 Dec	90
ANOM-DEC	Low	1-m SWE	1 Dec	240

^a Normal snow = 1 Dec 1985 initial conditions.

^b High snow = 1 Dec 1963 initial conditions with an avg initial snow cover anomaly of +6.7 mm SWE relative to normal.

^c Low snow = 1 Dec 1972 initial conditions with an avg initial snow cover anomaly of -5.0 mm SWE relative to normal.

while the other set was based on snow cover for November, December, and January as well as 1 December. The anomalies were obtained by subtracting each year's February (or December) monthly average snow cover from the long-term monthly mean February (or December) snow cover, and then normalized by dividing through by the standard deviation about the long-term mean. Figure 5 shows the time series of normalized anomalies for the February and December experiments. February and December were chosen for the same reasons as with the ANOM experiments, to assess the importance of time of year on the predictability of snow cover.

Each experiment set contains an ensemble of five runs. The purpose in making five-member ensembles for each case is to help evaluate the effects of internal variability within the model (a crude "signal-to-noise" measure) without compromising computational resources (five ensemble members give sufficiently robust results). One set had all runs begun on 1 February radiation date, and to obtain random but physically consistent perturbations, used as initial conditions the atmospheric states obtained for 30 and 31 January, and 1, 2, and 3 February. The other set had all runs begun on 1

December, with perturbation initial conditions obtained for 29 and 30 November, and 1, 2, and 3 December. The perturbation runs indicate the degree of inherent variability expected from the model, and represent a measure of "noise" above which any "signal" in snow cover must exceed. For simplicity, we will show the results of these ensemble sets as an ensemble mean and ± 1 standard deviation about this mean.

SPRED1 and SPRED3 both have initial atmospheric states from a "normal" February (1 February 1971), but forced with either high (1 February 1967; SPRED1) or low (1 February 1969, SPRED3) initial snow cover conditions. These experiments help assess the degree to which the subsequent extent of the snow cover depends solely on the initial state of the snow cover. SPRED2 and SPRED4 both have initial snow cover taken from the normal February of 1971 but initial atmospheric states for a high snow period (1 February 1967, SPRED2) or a low snow period (1 February 1969, SPRED4). A similar strategy is employed for the December runs (SPRED5-SPRED8), with the normal year taken from 1 December 1985 snow or atmospheric conditions, the high snow year taken from 1 December 1963 conditions, and the low snow year taken from 1 De-

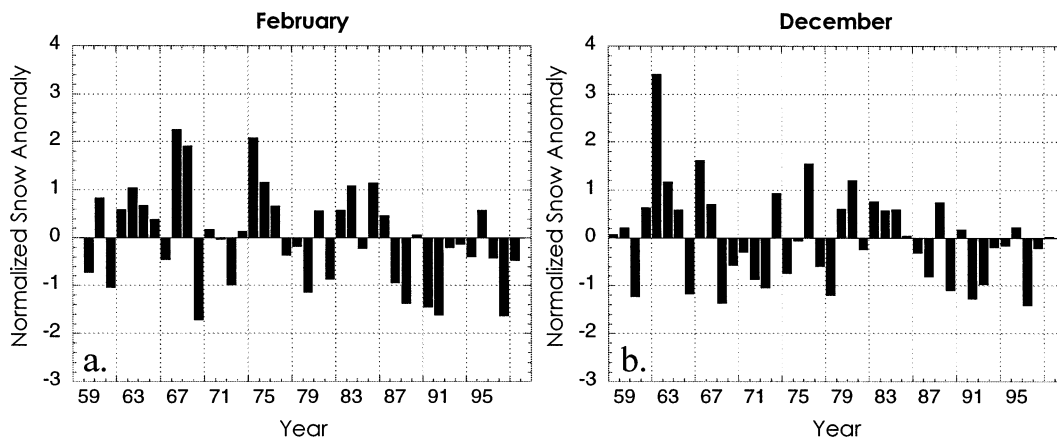


FIG. 5. Time series of normalized monthly snow cover anomalies over the western U.S. domain for (a) Feb and (b) Dec.

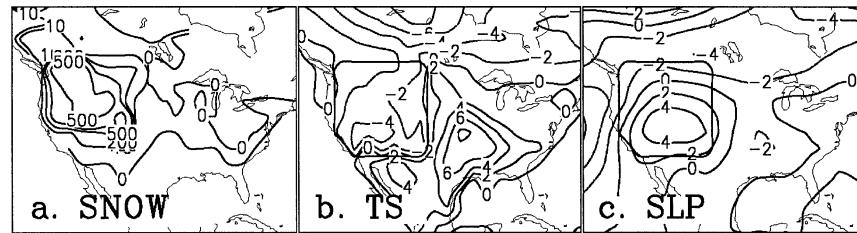


FIG. 6. ANOM-FEB minus CCM3/SST (low snow year of 1969) anomalies of (a) snow water equivalent (mm), (b) surface temperature (TS, °C), and (c) sea level pressure (SLP, mb) averaged for the month of Apr, following a start date of 1 Feb.

ember 1972 conditions (see Table 2 for description of December SPRED experiments). For both the February and December runs, the SSTs are selected from the same year as the initial atmospheric state. Thus, the only quantity we vary is the initial snow cover. These experiments assess the degree to which the high or low snow conditions depend solely on the initial state of the atmosphere. In each case we also impose the global snow cover state from these years as the initial anomaly to the perturbed run (this means that there are anomalies elsewhere on the globe). The years themselves were chosen based solely on the anomalies over the western U.S. region.

Figures 3a–c and 4a–c show the western U.S. domain that we use, as well as snow cover for December and February of those years in the CCM3/SST simulation used in the high and low SPRED experiments (see also Fig. 5). Figures 3d and 4d also show the anomalously large snow cover imposed for the ANOM runs. The western U.S. region is broadly defined as the region encompassing much of the Rocky Mountain and intermountain high plains.

5. Results and discussion

The presentation and discussion of the results focuses on: 1) The physical processes by which anomalous snow

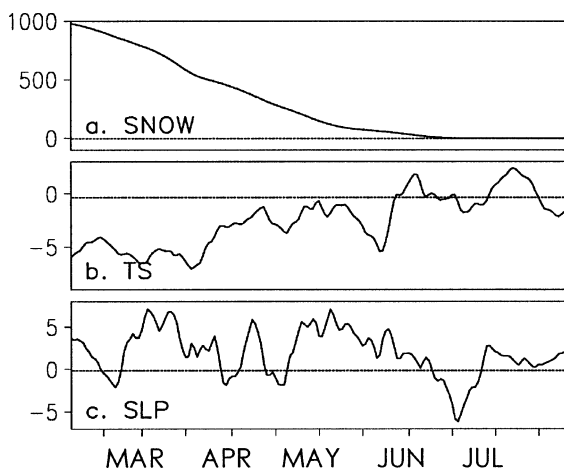


FIG. 7. Time series of ANOM-FEB minus CCM3/SST differences in (a) snow water equivalent (mm), (b) surface temperature (TS, K), and (c) sea level pressure (SLP, mb).

cover can be an active participant impacting the atmospheric state. 2) The ability of high (low) snow cover events to perpetuate, given atmospheric states from normal snow cover conditions during the accumulation season. 3) The degree of forcing that the initial atmospheric state from high (low) snow cover exerts on the subsequent evolution of an initial snow cover set to the mean state. 4) The extent to which time of year can be important, for example, early versus late winter.

a. ANOM experiments

1) FEBRUARY

In Fig. 6 are shown geographic plots of snow cover, surface temperature, and surface pressure, averaged over the month of April, for the ANOM-FEB minus CCM3/SST differences. Figure 7 shows a time series of the same differences, area-averaged over the western U.S. domain from February through July. The physical feedback effects between the snow and the atmosphere are shown clearly in these figures. An anomalously high snow cover (depth and extent) tends to maintain conditions favorable for the perpetuation of that snow cover. That is, high land surface albedo leads to decreased absorption of shortwave radiation at the surface, which, in turn, leads to cooler surface and near-surface air temperatures (especially along the snow edge) and decreased snowmelt (Figs. 6a,b, Figs. 7a,b). The cooler temperatures and decreased rate of snowmelt by themselves will tend to increase atmospheric stability; the cooling also leads to an increase in atmospheric pressure and hence changes in atmospheric circulation, which helps to maintain atmospheric stability (Figs. 6c and 7c). These changes in atmospheric circulation also manifest themselves downstream of our study region, as evidenced in Fig. 6 by warmer surface temperatures and reduced surface pressures over the central United States. The anomalous snow cover persists until early June, after which time surface temperature and surface pressure also rebound to near-normal values (Fig. 7).

2) DECEMBER

Figures 8 and 9 show the same fields for the December simulation. As in February, higher snow cover (Figs.

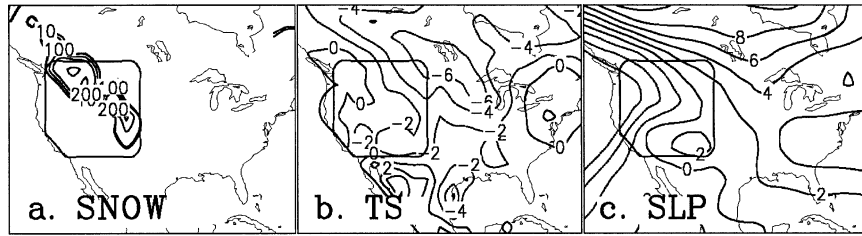


FIG. 8. ANOM-DEC minus CCM3/SST (low snow year of 1972) anomalies of (a) snow water equivalent (mm), (b) surface temperature (TS, °C), and (c) sea level pressure (SLP, mb) averaged for the month of Feb, following a start date of 1 Dec.

8a and 9a) results in higher surface albedo, which, in turn, leads to cooler surface temperatures (Figs. 8b and 9b) and increased atmospheric pressure (Figs. 8c and 9c). While the basic feedback mechanisms are the same as seen in the February case, the forcing on the atmosphere is not as persistent throughout the months following a December start date. By February, these effects are muted. This is partly due to the lessened strength of the albedo feedback during periods of low insolation, but also because the snow cover from the target “low” case (1972) had grown throughout the winter accumulation season. The difference in snow cover between the ANOM-DEC experiment and the (initially) low snow cover year had essentially disappeared by early March (Fig. 9a). Therefore, the anomalous forcing on the atmosphere did not persist into the spring melt season. Both the low insolation and the accumulation of snow cover during the winter season suggest that the time of year does play a role in the strength of the feedback of the snow cover. We might expect to see these same seasonal differences (although more muted) in the predictability experiments.

These results with a large imposed initial snow cover also have implications for studies of ice sheet inception (e.g., Oglesby 1990; Oglesby and Marshall 1997; Mar-

shall et al. 1999). A key element of these earlier studies was the imposition of an initial, very large snow anomaly over large regions, including those thought to be important for the start of ice sheets. The new results presented here define the basic physical processes and suggest how even in this midlatitude region, an anomalous snow cover can persist for months and condition its environment to help that perpetuation.

Figure 10 shows time series averaged over the western U.S. region for differences in planetary (top of the atmosphere) albedo, surface albedo, and fractional total cloudiness between the 1-m anomaly cases and the control run. It is clear that for the anomaly run started in February (Figs. 10a–c), the positive anomalies in planetary albedo (Fig. 10a) through April are largely due to the positive anomalies in surface albedo (Fig. 10b), which, in turn, result from the enhanced snow cover. Those days that have strong negative anomalies in total cloudiness (Fig. 10c) serve only to bring the planetary albedo anomaly close to zero, or at most, slightly negative. Furthermore, positive and negative anomalies in total cloudiness essentially cancel out over this period. Only after the anomalous snow has largely melted by the end of April does surface albedo become relatively constant, with anomalies in planetary albedo corresponding to anomalies in total cloudiness. Results for the anomaly run started in December (Figs. 10d–f) are similar, stressing even in this run the importance of the albedo effect of snow. The anomalies in surface and planetary albedo (Figs. 10e and 10d, respectively) are as large in the December run as in February, but the absolute amount of solar radiation during this low sun time of year is simply insufficient to translate the albedo effects into a strong climatic effect.

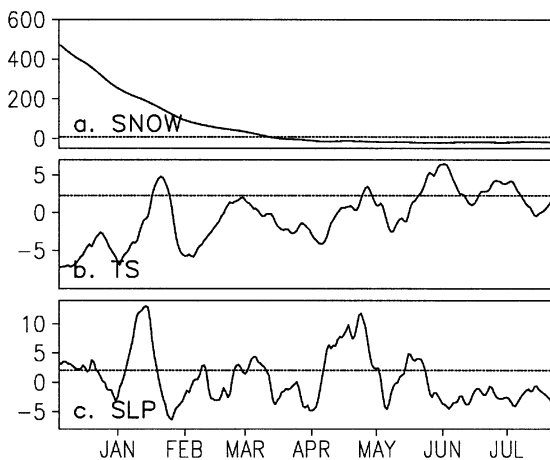


FIG. 9. Time series of ANOM-DEC minus CCM3/SST differences in (a) snow water equivalent (mm), (b) surface temperature (TS, K), and (c) sea level pressure (SLP, mb).

b. SPRED experiments

1) FEBRUARY

(i) High snow

Figure 11 shows the results from the (a) SPRED1 and (b) SPRED2 simulations. This figure shows the ensemble mean SWE of the SPRED experiments (solid gray curve) as well as the spread in the ensembles (± 1 standard deviation, shown as a dashed gray curves). The

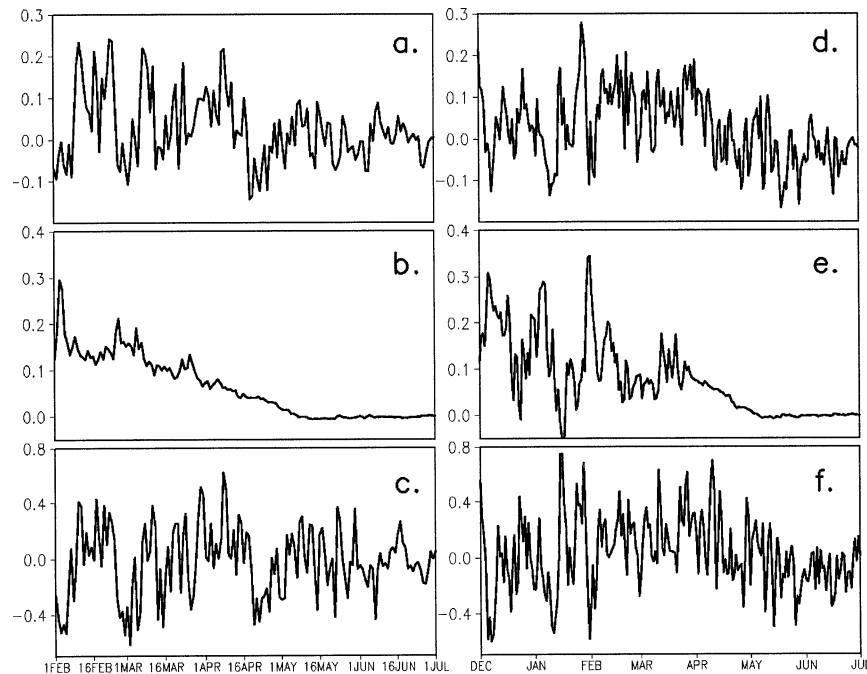


FIG. 10. Time series of ANOM minus CCM3/SST anomalies averaged over the western U.S. region. Anomalies from the ANOM-FEB experiments for (a) planetary albedo, (b) surface albedo, and (c) total cloudiness, and anomalies from the ANOM-DEC experiments for (d) planetary albedo, (e) surface albedo, and (f) total cloudiness.

ensemble statistics are compared to two results from the CCM3/SST simulation: (1) the “normal” year SWE (in this case, 1971) and (2) the “high” year SWE (1967). It is obvious that the SPRED1 ensembles maintain the extensive initial snowpack for at least 6 weeks, despite an initial state of the atmosphere taken from the normal

snow year of 1971 (Fig. 11a). The SPRED2 ensembles (Fig. 11b), on the other hand, maintain the much lower initial snow cover of 1971, despite an initial atmospheric state taken from the high snow year of 1967.

(ii) Low snow

Figure 12 shows the results from the (a) SPRED3 and (b) SPRED4 simulations. The SPRED ensemble mean (solid gray curve) and ± 1 standard deviation (dashed gray curves) are displayed as well as the normal and low anomaly from the CCM3/SST simulation. The SPRED3 ensembles clearly maintain the low initial snow cover of 1 February 1969 despite a mean initial state of the atmosphere. The SPRED4 ensembles maintain a much higher degree of snow cover than in the base 1969 case, despite an atmospheric initial state consistent with low snow conditions.

The preceding figures indicate the robustness of these results. In both sets of simulations (high and low snow cases) the ensembles track the climate of the initial snow cover for several weeks. These analyses suggest that for runs starting in February, the initial snow cover is much more important than the initial state of the atmosphere in predicting the extent of snow cover for the next several months. For the high snow anomalies this results, in part, from the increased surface albedo which leads to cooler temperatures. This process is seen in our results for the ANOM-FEB case (see Figs. 6 and 7) and is also evident in the results for the SPRED1 and

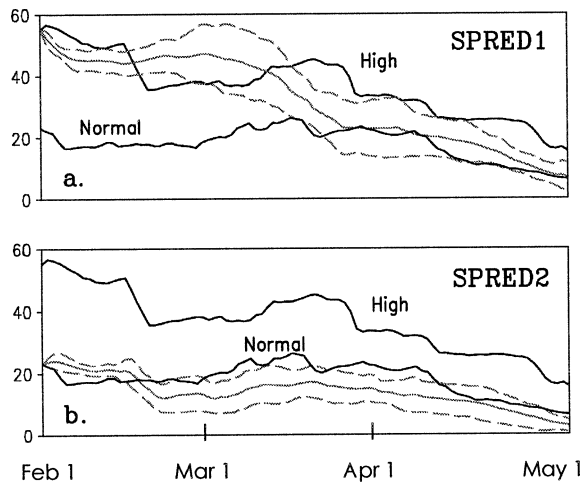


FIG. 11. Time series of (a) SPRED1 (high initial snow cover, normal snow atmosphere) and (b) SPRED2 (normal initial snow cover, high snow atmosphere) ensemble average snow water equivalent (mm) compared to the control high snow year (1967) and normal snow year (1971). Gray lines represent the mean (solid line) and standard deviation about the mean (dashed line) of the SPRED ensembles.

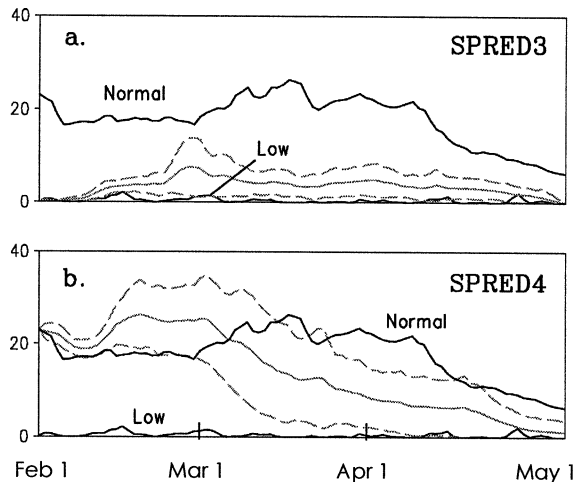


FIG. 12. Time series of (a) SPRED3 (low initial snow cover, normal snow atmosphere) and (b) SPRED4 (normal initial snow cover, low snow atmosphere) ensemble mean snow water equivalent (mm) compared to the control low snow year (1969) and normal snow year (1971). Gray lines represent the mean (solid line) and standard deviation about the mean (dashed line) of the SPRED ensembles.

SPRED2 cases. A complementary mechanism occurs via the energy required to melt the snowpack. With the presence of snow cover over the land surface, any excess energy would go towards raising the snowpack temperature to 0°C and the melt/sublimation of the snowpack. This energy is then unavailable to heat the land surface. On the other hand, for low snow anomalies, a decreased snow extent would allow for increased absorption of shortwave radiation at the surface leading to warmer surface temperatures. Subsequently, in high snow years the overlying atmosphere tends to remain cooler while in low snow cover years such cooling effects are reduced or even absent.

2) DECEMBER

Figure 13 shows the results from the (a) SPRED5 and (b) SPRED6 experiments. Results from the SPRED7 and SPRED8 experiments (not shown) have similar interpretation and implications. The December results are less clear than for February. The high SWE of the SPRED5 (and CCM3/SST) simulations are maintained for several weeks but then become indistinguishable from the normal year snow cover as the snow cover accumulates in the winter. Similarly, the normal snow conditions of SPRED6 are not significantly different from the high snow conditions of CCM3/SST after a few days. The expected accumulation of snow cover during this season overshadows any effect the snow cover or atmospheric state may have on the subsequent snow accumulation. A similar result was seen in the DEC-ANOM experiment when an unrealistically large snow anomaly was imposed over the western United States at the beginning of the accumulation season. The snow cover albedo feedback is also less active during

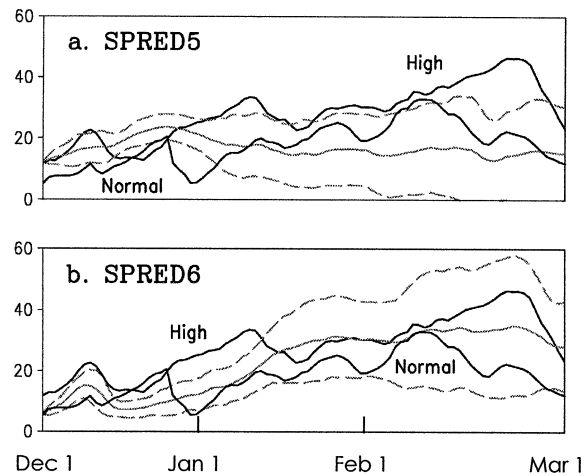


FIG. 13. Time series of (a) SPRED5 (high initial snow cover, normal snow atmosphere) and (b) SPRED6 (normal initial snow cover, high snow atmosphere) ensemble average snow water equivalent (mm) compared to the control high snow year (1963) and normal snow year (1985). Gray lines represent the mean (solid line) and standard deviation about the mean (dashed line) of the SPRED ensembles.

this time due to the lower insolation. It is suggested here that, during the early part of the snow accumulation season, the initial snow cover state is less important than in the spring. However, the SPRED5 results do suggest that the initial snow cover state is more important than the initial atmospheric state in maintaining snow cover.

c. Discussion

Overall, we find that initial snow cover anomalies in February have a significantly greater impact on ensuing snow depth and surface temperature than do snow anomalies imposed in December. One key explanation for the seasonal differences is in the effect of the snow albedo on the winter climate. During the winter accumulation season, the albedo effect is lessened by the low insolation, but later in the season (and into the spring ablation season), the albedo effect becomes stronger with higher insolation. A competing effect during the early winter is the subsequent accumulation of snow cover, which acts to mute any initial differences. An initially low snow cover in early December does not appear to impede subsequent accumulations of snow cover later in the winter season.

Another factor that enhances the predictability of late winter/early spring snow cover anomalies is the energy required for snowmelt. The albedo effect, because it causes more reflection of incoming shortwave radiation, results in less energy available at the surface for heating and melting the snowpack. This also keeps the surface cooler and, in turn, helps to maintain a cold snowpack. As noted earlier, in the ANOM-FEB runs, a significant fraction of the surface energy balance goes to latent heat of snowmelt and is not otherwise available for surface heating. Complementary to this effort, in related work

we performed a study looking at the predictability of soil moisture analogous to the snow study presented here (Oglesby et al. 2002, hereafter OMRR). In contrast to our snow cover results, in the soil moisture runs, the initial state of the atmosphere appears more important than the initial state of soil moisture. Since over 7 times as much energy is required to evaporate a given mass of liquid water than to melt the same mass of frozen water, the implication is that the energy required to melt snow cover may *not*, however, be the major cause of the dependency on initial snow cover state. Rather, the snow albedo effect described earlier appears to be the dominant mechanism.

6. Conclusions

We have made a series of snow cover predictability runs with the NCAR/CCM3 based on anomalous snow cover over the western United States. An existing 42-yr simulation forced by observed SST was used as our baseline. Two basic types of simulations were made, starting on 1 December (early in the snow season) and 1 February (late in the snow season): 1) Simulations with exaggerated initial snow cover of 1-m water equivalent over the western United States (called ANOM) and 2) predictability experiments in which snow and atmospheric states from different years were combined (called SPRED).

The ANOM simulations were undertaken in order to elucidate the physical mechanisms responsible for the snow-atmosphere feedbacks. These runs made it clear that the high albedo of snow was a key physical process; in February, when the sun is relatively high in the sky and days are longer, the albedo effect became the dominant factor; while in December the sun was too low in the sky and days too short for the albedo effect to dominate. Energy required to melt the snow cover also played a role, although comparison to the analogous soil moisture predictability study of OMRR suggests that it is a secondary factor. The second type of simulations (SPRED) were undertaken to assess the degree of predictability in the system and to assess the relative importance of the initial snow cover versus the initial state of the atmosphere in determining the subsequent evolution of the snow cover. These experiments included two kinds of simulations for each time of year: 1) Simulations in which the initial snow cover was taken from an anomalously high or low snow cover year, and the initial state of the atmosphere from a year with “near-normal” snow cover. 2) Simulations in which the initial state of the atmosphere was taken from a year with anomalously high or low snow cover, while the initial snow cover was taken from a near-normal year. The results of these experiments showed that the initial snow cover was more important than the initial state of the atmosphere in determining the subsequent state of the snow cover (and hence the surface climate). With the inherent natural variability in the climate system, the

timescale of predictability was anywhere from 2 to 6 weeks.

These results indicate that the time of year can be crucial. When introduced in late winter, the anomalies strongly affected the subsequent evolution of snow cover. This was evident in both the SPRED and ANOM experiments. However, when introduced in early winter, much less effect is seen on the subsequent snow cover in the SPRED experiments and a more short-lived effect is seen in the ANOM experiments. Runs with greatly exaggerated initial snow cover indicate that the high reflectivity of snow is the most important process by which snow cover can impact climate, through lower surface temperatures and increased surface pressures. In early winter, the amount of solar radiation is very small and so this albedo effect is inconsequential; while in late winter, with the sun higher in the sky and period of daylight longer, the effect is much stronger. The subsequent accumulation of snow in the December ANOM run as the winter season progressed also helped to mask the effects of the initial anomaly.

The fact that some predictability appears inherent in the snow cover climate (particularly in modifying the atmospheric circulation) suggests that these results may have implications toward downstream and interseasonal effects on the climate outside of the snow-covered area. Such effects have been suggested by several studies (Gutzler and Preston 1997; Smith and O’Brien 2001; Kunkel and Angel 1999; Janowiak and Bell 1998). These studies suggest a possible connection between the El Niño–Southern Oscillation (ENSO), anomalies in North American snow cover, and the North American Monsoon System (NAMS). Although this study does not directly address the summer monsoon over the southwestern United States, the CCM3/SST simulations and the ANOM experiments provide a template for looking for a possible climate signal. Future work includes model experimentation to examine the role of late-season (i.e., ablation season) snow cover in modulating early-summer precipitation anomalies and the summer monsoon in the southwestern United States.

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