A Model Study of the Short-Term Climatic and Hydrologic Effects of Sudden Snow-Cover Removal

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ABSTRACT

This paper describes the results from a set of numerical experiments which stimulate the effect of a large-scale removal of snow cover in middle and high latitudes during the early spring season. This is done through use of a simplified general circulation model with a limited computational domain and idealized geography.

It is found that removal of snow cover reduces the water available to the soil through snowmelt and decreases soil moisture in this region during the following seasons. Furthermore, it also reduces surface albedo in this region and increases absorption of insolation by the ground surface. This, in turn, heats the ground surface and allows more evaporation to occur. However, the change of evaporation is relatively small owing to the low values of surface temperature in high latitudes. Therefore, the negative anomaly of soil moisture induced initially by the removal of snow cover persists for the entire spring and summer seasons.

The removal of snow cover also affects the thermal and dynamical structure of the atmosphere. It is found that the increase of surface temperature extends into the upper troposphere thereby reducing both meridional temperature gradient and zonal wind in high latitudes.

1. Introduction

It has long been noticed that the albedo of large-scale snow and ice cover is an important climatic factor, but in the past this factor has been considered in relation to studies of long-term averages of climate. In the last two decades, work has also been done in determining its effect on short-term climatic change. Namias (1963) studied the feedback effect of large-scale winter snow on the atmospheric circulation. Hahn and Shukla (1976) investigated the possibility that a significant relationship exists between winter Eurasian snow cover and monsoon rainfall in India. Chen and Yan (1978) showed that the summer circulation over the Tibetan Plateau and its vicinity arrived later if the winter snow cover over the plateau was great and earlier if snow cover was small.

These studies show that large-scale snow cover not only has an immediate influence on atmospheric circulation but it can also have a long-lasting effect on climate. The reason for its long lasting effect may be visualized as follows. During its existence, the snow cover can increase the surface albedo considerably. The large-scale decrease of ground absorption of solar radiation due to the increase of surface albedo can produce a substantial influence on the heat budget of both earth's surface and atmosphere, and accordingly, on the atmospheric circulation (Charney, 1975). The contemporary change of circulation will also cause a change of the ground water budget, for instance, through changes in precipitation. The water of the melting snow will penetrate into the soil, increasing the content of soil moisture. This increase will, on one hand, decrease the albedo of the soil (although very slightly) and, on the other hand, will enhance the ground surface evaporation. These changes will produce variations of the water budget and heat budget near the ground and will have a feedback effect on atmospheric circulations and hydrological processes.

In some winters the snow cover is very widespread in high and middle latitudes and in other winters much less snow cover is observed. It may be expected from the above that the large-scale variation of the snow cover will cause differences between the two types of years in the climate and hydrology of later time periods. The present paper attempts to simulate these differences through integration of a GCM with surface hydrology and an oceanic layer. Through this simulation the mutual link between atmospheric and hydrological processes may also be studied.

2. Description of the model

To investigate the influence of large-scale snow cover on short-term climatic and hydrological change, we used the sector model described by Wetherald and Manabe (1981). It is similar to the model developed by Manabe and Stouffer (1980) except that it has a limited computational domain and idealized geography illustrated in Fig. 1. It has three basic units: 1) a general circulation model of the atmosphere, 2) a

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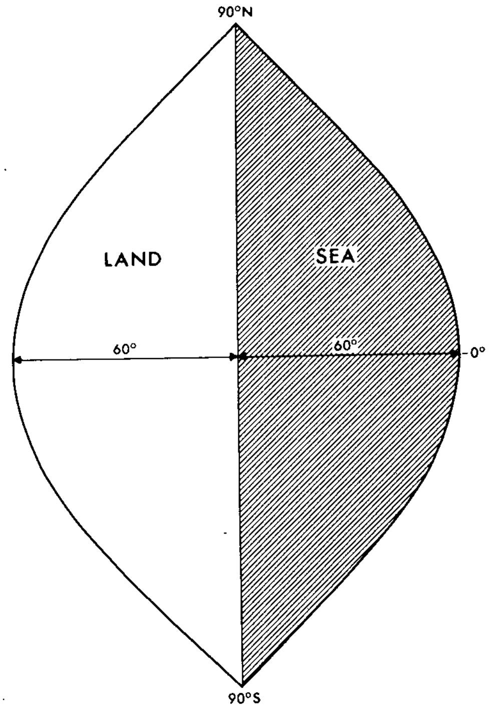

FIG. 1. Diagram of the computational domain of the model.

heat and water balance model over the continents, and 3) a simple model of the mixed layer in the oceans. The atmospheric model computes the changes of vertical component of vorticity, horizontal divergence, temperature, moisture and surface pressure. The horizontal distributions of the atmospheric variables are represented by spherical harmonics. Fifteen orthogonal components are retained in both zonal and meridional directions. The vertical derivatives appearing in the prognostic equations are computed by finite difference method from the variables specified at nine unevenly spaced levels.

The distribution of incoming solar radiation is prescribed at the top of the atmosphere with a seasonal variation but with no diurnal variation. Both the solar and terrestrial radiation are affected by carbon dioxide, ozone, water vapor and cloud cover. The mixing ratio of carbon dioxide is assumed to be constant everywhere and ozone is specified as a function of latitude, height and season. The prescribed distribution of cloud cover is zonally uniform and invariant with season. The water vapor is determined prognostically. Whenever supersaturation occurs precipitation is predicted in the form of rain or snow according to whether air temperature near the surface is above or below freezing respectively. The moist convection processes are incorporated by a moist convective adjustment scheme (Manabe *et al.*, 1965).

Over the continent, the earth's surface temperature is determined by the surface heat budget. The soil albedo, required for computation of the solar flux at the ground, is prescribed zonally (see Manabe, 1969) but replaced by a value of 0.60 over the region covered with snow at a temperature higher than −10°C and by a value of 0.70 over the region covered with snow at a temperature lower than −10°C. A change of snow depth is calculated from the net contribution of snowfall, sublimation and snowmelt.

The ground water budget is computed as follows. It is assumed that the field capacity of the soil is 15 cm and has no geographical variation. The change of soil moisture is determined from rates of rainfall, evaporation, snowmelt and runoff. Runoff is predicted when the soil moisture exceeds 15 cm. The rate of evaporation from soil is a function of soil moisture and the hypothetical rate of evaporation from completely saturated soil. Sublimation is computed in place of evaporation in regions where snow cover exists. In this event, moisture loss takes place from the snow surface rather than from the underlying soil surface.

The ocean is simply a mixed isothermal layer with a uniform depth of 68.5 m. The rate of change of temperature in ice-free regions is computed from solar and terrestrial radiation fluxes and sensible and latent heat fluxes at the ocean surface with the effects of ocean currents and the heat exchange between the mixed layer and deeper layer of the ocean being neglected. In the presence of sea-ice, the mixed layer temperature is fixed at −2°C, the freezing point of sea water, and the heat conduction through the ice is balanced by latent heat of freezing (or melting) at the bottom of the ice layer. The ice thickness is determined by this process together with the melting at the upper ice surface, sublimation and snowfall (Bryan, 1969). Albedo at the oceanic surface is prescribed as a function of latitude, but a high value of albedo is used over the region covered by sea ice. It is 0.70 if the temperature of the sea ice surface is below −10°C and 0.60 if it is above this value.

3. Plan of experiment

As was stated in Wetherald and Manabe (1981) the results for the standard experiment were averaged over the last four years of the entire integration. However, since the two hemispheres are identical, this implies an effective eight-year average if the two hemispheres are averaged together after shifting the phase of seasonal variation in the Southern Hemisphere by a half year. Therefore, the experimental procedure for the current study consists of setting the snow depth to zero initially for eight separate runs; four for the "Northern Hemisphere" and four for the "Southern Hemisphere." Hereafter, this set of new runs will be referred to simply as RS (removal of snow). The initial date chosen for the Northern Hemisphere runs is 15 March whereas the corresponding date for the Southern Hemisphere runs is 15 September. Both sets are averaged together to yield an eight-year average starting at the effective date of 15 March for convenience. All eight runs are integrated for a period of more than five months. (Note that after the initial date, snow may still accumulate on the ground surface in the model.) Thus, analysis of the influence of large-scale snow cover will be based mainly on the difference between the RS and standard experiments.

To evaluate the performance of the basic model, we show the winter and summer distribution of precipitation rate obtained from the standard experiment (Fig. 2a, 2b). (This figure is reproduced from the study by Manabe *et al.* (1981), hereafter referred to as MWS.) For the winter season (Fig. 2a), one may identify an oceanic tropical rain belt, a subtropical dry zone and the mid-latitude rainbelt centered at about 45° latitude. In addition, one may identify a region of relatively high precipitation rate along the east coast of the subtropical portion of the idealized continent as compared with a low precipitation rate along the west coast. Turning to the summer situation (Fig. 2b), one notes that the tropical rainbelt is considerably more intense and, similar to a monsoon pattern, extends westward well into the continent. Along with the poleward shift of the mid-latitude rainbelt, the subtropical dry zone has also extended northward. From the preceeding we can see that, although this model is idealized, its stimulation is qualitatively realistic with regard to the main hydrologic features.

The amount of snow which is artifically removed in the RS experiment is illustrated by Fig. 3a which shows the geographical distribution of snow depth in water equivalent over the idealized continent on the date of snow removal (eight-year average). This figure indicates that the snow line is near 49° latitude, and a belt of zonal maximum depth exists around 70° latitude and a region of local maximum near 60° longitude (i.e., the east coast of the idealized continent in the polar region). After removal of this snow cover, the distribution of soil moisture content will play an important role in the forthcoming atmospheric and hydrological processes of the perturbed experiments. Fig. 3b illustrates the geographical distribution of soil moisture on the date of snow removal. (See also Fig. 7b of MWS for the seasonal variation of zonal mean soil moisture from the standard experiment.)

It is our opinion that the proposed experiment has some relevance to a real situation. For example, during a certain spring season the extent and depth of snow cover can be significantly below normal because of a low snowfall rate in winter or rapid snowmelt in early spring. This study attempts to evaluate the consequence of a below-normal snow cover in spring by comparing the results from the RS-experiment with those from the standard experiment.

4. The change of heat balance near the ground

As stated above, after the snow cover is removed on 15 March, snow may still fall and accumulate on the ground. The latitude-time distribution of zonal mean difference of snow depth between the RS experiments and the standard experiments is shown in Fig. 4.

The most direct effect of the removal of snow cover is to decrease surface albedo. This, in turn, increases

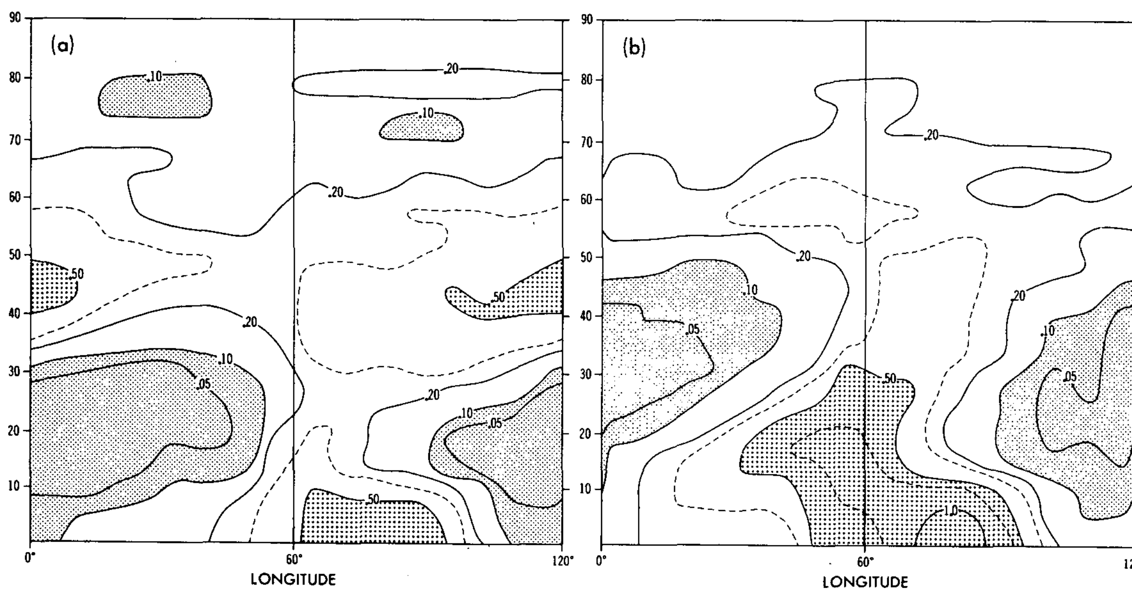

FIG. 2. Geographical distribution of the 3-month mean precipitation rate obtained from the standard experiment (as in Fig. 6 of MWS) for (a) the winter season and (b) the summer season. The distributions of the two hemispheres of the model are averaged after shifting the phase of the Southern Hemisphere variation by six months. Units are in cm day^{-1}.

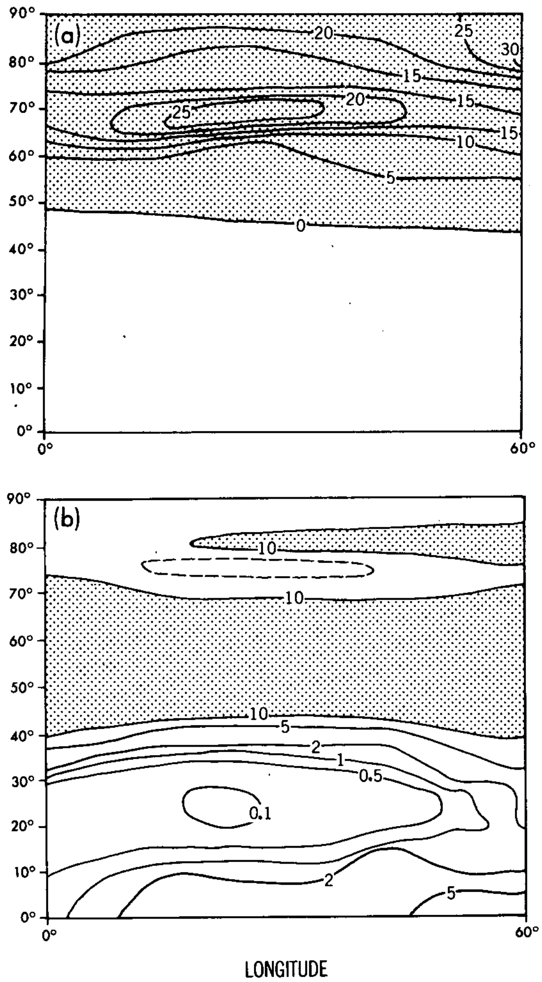

FIG. 3. Geographical distribution of (a) initial snow depth in water equivalent over the continent (eight-year average) and (b) initial soil moisture over the continent (eight-year average). Units are in cm.

absorption of solar radiation in this latitude zone which changes the heat balance of the ground considerably. The decrease of surface albedo from the standard for the average of the eight runs is given in Fig. 5. This figure gives the latitude–time distribution of the zonal mean difference of surface albedo (over land) between the RS experiments and the standard experiments. Since snow cover is removed on 15 March of the model year, the March snow depth is not shown here. The same is true for the month of September since the calculation is terminated on 15 September. (The results from the four Southern Hemisphere runs are also incorporated after shifting of the phase by six months.) This figure is quite representative for each of the eight separate runs because variation from one run to another is quite small. From this figure, it can be seen that in the region where the snow cover is removed, there is a significant reduction of surface albedo from April until July. Near the pole, the reduction is at a maximum and approaches 20% in May. In April, the maximum decrease occurs in a latitude band near 70°. As time proceeds, this maximum shifts northward and reaches the pole by May. This is simply a result of the snow cover melting at increasingly higher latitudes with time as the warm season approaches.

Before passing on to a discussion of the heat balance of the continental surface of the model, it is worthwhile to briefly describe the relevant equation which describes this process.

The heat balance requirement over the continental surface may be written as

$$[SW] - [LW] - [SM] - [SH] - [LH] = 0, \quad (1)$$

where the brackets indicate a zonal mean value, SW is the net solar radiation, LW net long wave radiation, SM snowmelt, SH sensible heat and LH is latent heat. Here SW is defined as positive downward, whereas LW, SH, and LH are considered positive upward.

The difference of the heat balance components between the RS and standard experiments may be given by the following relationship which is derived from Eq. (1)

$$\Delta[SW] - \Delta[LW] - \Delta[SM] - \Delta[SH] - \Delta[LH] = 0, \quad (2)$$

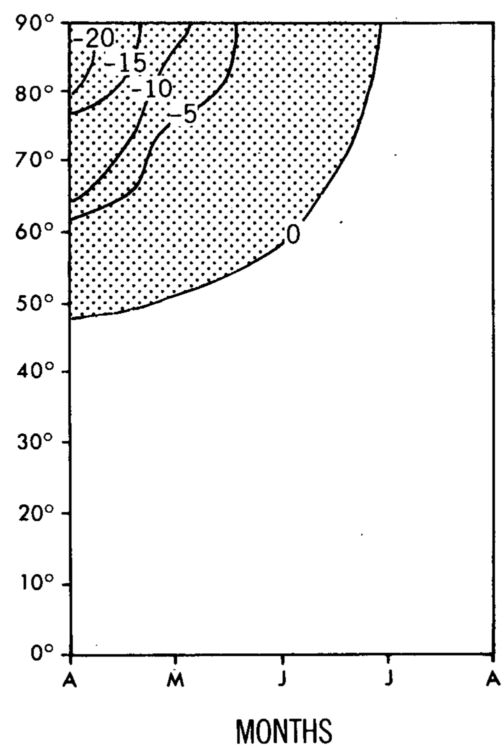

FIG. 4. Latitude–time distribution of the seasonal variation of the zonal mean difference of snow depth over the continent between the RS and standard experiments for the months of April through August. Units are in cm of water equivalent.

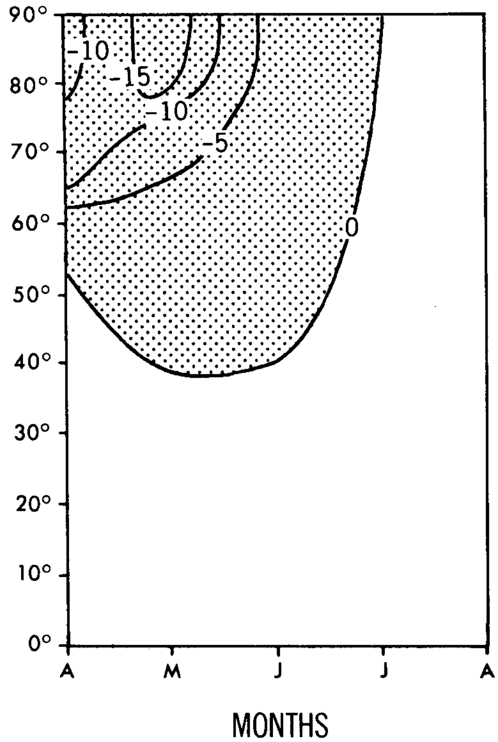

FIG. 5. Latitude–time distribution of zonal mean difference of surface albedo between the RS and the standard experiment over the continent. Units are in %.

where Δ denotes the difference between the two experiments (RS minus standard).

A direct consequence of decreasing surface albedo is the increase of absorption of solar radiation by the ground which affects the heat balance near the ground in the following months. Fig. 6 illustrates this process. In this figure, the variation of zonally-averaged monthly-mean difference between the RS and standard experiments of various surface heat balance components at 75° latitude over the continent are shown. This latitude is chosen because of its location at the center of the region of snow cover removal. All of the five curves show an extreme value in May which results from the maximum decrease of surface albedo in this month at 75°N. From the curve for the change of insolation ($\Delta[SW]$) one can see that during the entire spring and summer, $\Delta[SW]$ is positive with a maximum value of 20 W m^{-2} in May. For evaporation, the removal of snow cover results in a large increase during April and May as indicated by the negative value of $-\Delta[LH]$. The increase of evaporation in these two months (more than 16 W m^{-2} in May) is large enough to cause considerable depletion of soil moisture during this period as discussed in the following section (see Fig. 9 and 10). This, in turn, causes the evaporation difference to reverse sign (become negative) from June onward as Fig. 6 indicates. For snowmelt, the curve shows a large decrease due to the removal of snow cover as indicated by the positive value of $-\Delta[SM]$. After June, the difference between the two experiments is

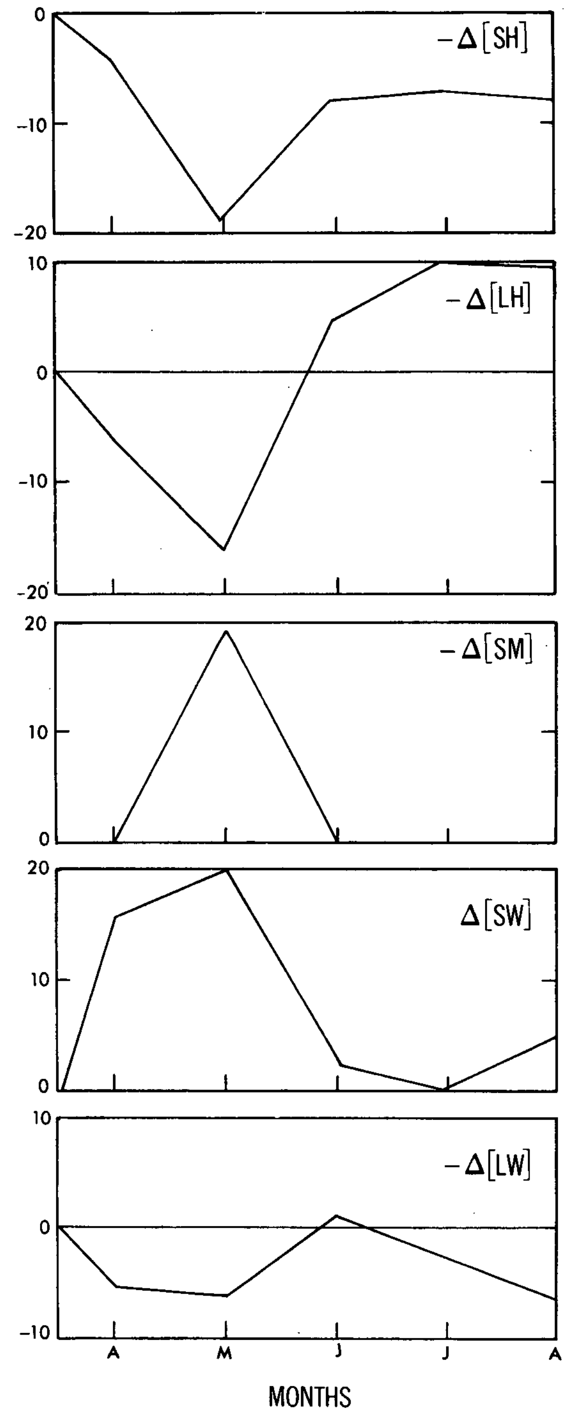

FIG. 6. Time variations of the zonally-averaged monthly mean differences in the heat balance components at 75°N latitude over the continent. The differences for March are set equal to zero. Units are W m^{-2}.

zero because by this time the snow cover has completely disappeared in both experiments.

Notice that the difference of sensible heat flux between the RS and standard experiments is positive (i.e., $-\Delta[SH]$ is negative) during the entire spring and summer seasons with the maximum in May amounting to 19 W m^{-2}. Again, this is because the surface absorption of insolation is enhanced and the latent heat for melting is reduced due to the removal of snow cover. The last curve of Fig. 6 to be discussed is the change of net long wave radiation from the ground. This curve is the least varied among the five quantities presented. The change of this quantity resulting from the removal of the snow cover is mostly positive (as indicated by the negative values of $-\Delta[LW]$) except for a slightly negative value in June. The increase of the long wave emission from the ground to the atmosphere is due to an increase of surface temperature, a topic to be discussed next.

The increase of surface temperature of the soil due to the removal of snow cover may be analyzed from changes of the five heat balance components shown in Fig. 6. From Fig. 6, it may be seen that $\Delta[SW]$ amounts to 15 W m^{-2} in April and 20 W m^{-2} in May due to the decrease of surface albedo. This will cause the soil temperature to increase. Fig. 7 illustrates the monthly variation of zonal mean difference of surface temperature over the continent at 75°N between the RS and standard experiments. As seen from this figure, the change of zonal mean surface temperature due to the removal of snow cover is positive from April until August with double maxima; one in May amounting to nearly 4°C and the other in July of more than 3°C.

Since the surface temperature of soil is an important and practical element in agriculture, we will discuss it in some detail. Fig. 8 illustrates the latitude-time distribution of zonal mean difference of surface temperature over the continent between the RS and standard experiments. As seen from this figure, the difference is positive almost everywhere in the region of snow cover removal. The increase of surface temperature is at a maximum (over 8°C) in May near the pole and decreases southward. The area of a 2°C or greater increase covers most of the entire region from April to July.

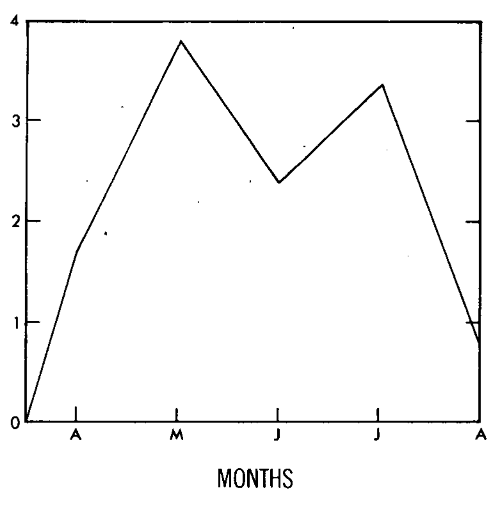

FIG. 7. Time variations of monthly zonal mean surface ground temperature difference at 75°N between the RS and standard experiments over the continent. Units are °C.

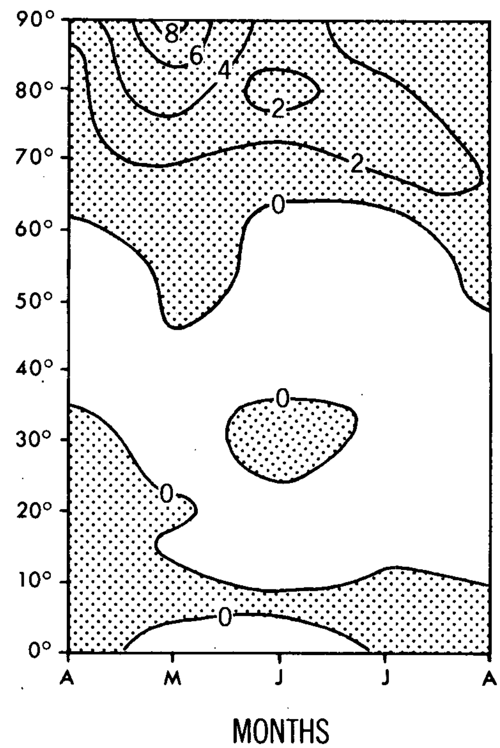

FIG. 8. Latitude–time distribution of zonal mean difference of surface ground temperature between the RS and standard experiments over the continent. Units are °C.

The above processes may be summarized as follows. Removal of snow will initially decrease the ground albedo which will, in turn, increase the absorption of solar radiation by the ground surface. Furthermore, it reduces the amount of heat required for melting snow cover. The additional heat, which is thus available, raises the temperature of the continental surface thereby increasing the upward fluxes of sensible heat, latent heat and terrestrial radiation during late spring and early summer.

5. The change of ground water balance

In the previous section, we have discussed the effect of the removal of snowcover on the heat balance near the ground. This will ultimately lead to changes of the ground water balance. To illustrate this, we first show the latitude-time distribution of the difference of the zonal mean soil moisture content between the

RS and standard experiments (Fig. 9). According to this figure, the difference of zonal mean soil moisture is negative everywhere in the region of snowcover removal from April to August. In other words, the area of negative soil moisture difference extends from 65°N to the pole for the entire spring and summer seasons. The maximum value of the difference is more than −4 cm near 75°N during the month of May.

To further study the reasons for this change of soil moisture, the soil moisture budget from both RS and standard experiments is examined. The prognostic equation of soil moisture may be written as (MWS).

$$\frac{\partial[w]}{\partial t} = [r] - [e] + [m] - [f], \tag{3}$$

where w is soil moisture, r the rate of rainfall, e the rate of evaporation, m the rate of snowmelt, f the rate of runoff and the brackets denote a zonal average over the continent.

The difference in zonal mean soil moisture budget between the RS and the standard experiments may be studied by the following relationship which is obtained from equation (3).

$$\frac{\partial\Delta[w]}{\partial t} = \Delta[r] - \Delta[e] + \Delta[m] - \Delta[f], \tag{4}$$

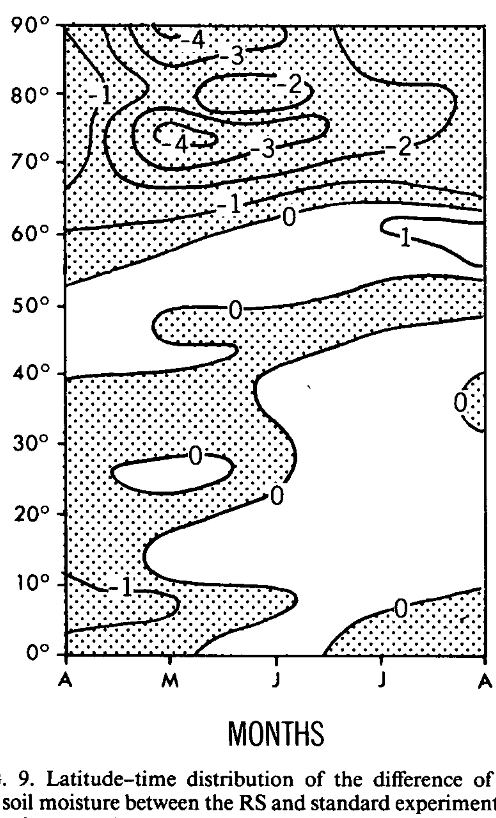

FIG. 9. Latitude–time distribution of the difference of zonal mean soil moisture between the RS and standard experiments over the continent. Units are in cm.

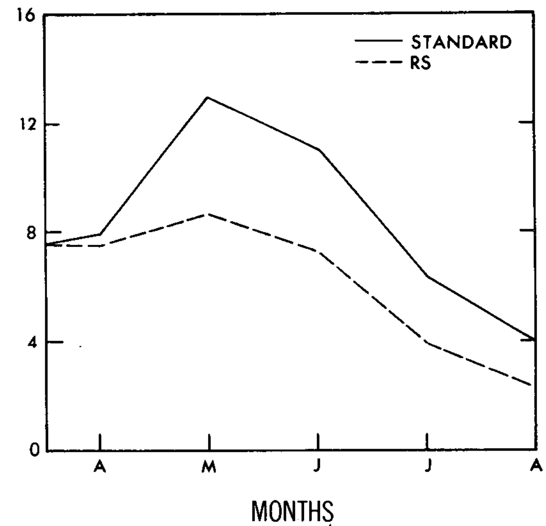

FIG. 10. Time variations of monthly mean soil moisture at 75°N of the RS experiment (dashed curve) and the standard experiment (solid curve) over the continent. Units are in cm.

where Δ denotes the difference between the two experiments.

In order to appreciate the relative magnitude of the above soil moisture change, we prepared Fig. 10 which gives the monthly variation of zonal mean soil moisture content at 75°N of the RS experiment (dashed curve) and the standard experiment (solid curve). The dashed curve is below the solid line everywhere, i.e. the soil moisture is depleted by the removal of snowcover. The difference between these two curves becomes a maximum in May and decreases from May until July. It is of interest to note that in Fig. 10 there is still a substantial negative difference of soil moisture between the RS and standard experiments by the end of the summer (August). This feature illustrates persistence of the negative soil moisture anomaly initially induced by the removal of snow cover.

To see how the variation of the difference between the two curves, i.e. $\Delta[w]$, is produced, we prepared Fig. 11 which gives the monthly variation of $\Delta[r]$, $-\Delta[e]$, $\Delta[m]$ and $-\Delta[f]$ at 75°N. The sum of these four quantities is equal to $\partial\Delta[w]/\partial t$ which is also given in Fig. 11. We will first consider the curve $\partial\Delta[w]/\partial t$. It shows a negative maximum in May (−0.113 cm day^{-1}). This negative maximum decreases from May to summer. It is easily seen by comparing the other four curves that this maximum comes mainly from the difference of $\Delta[m]$ and $\Delta[f]$. There is a large decrease of snowmelt from the RS experiment to the standard experiment due to the removal of snowcover. This is, of course, very reasonable since the snowcover was simply removed rather than allowed to melt prematurely in the RS experiment. Since the changes of rainfall amount and evaporation between

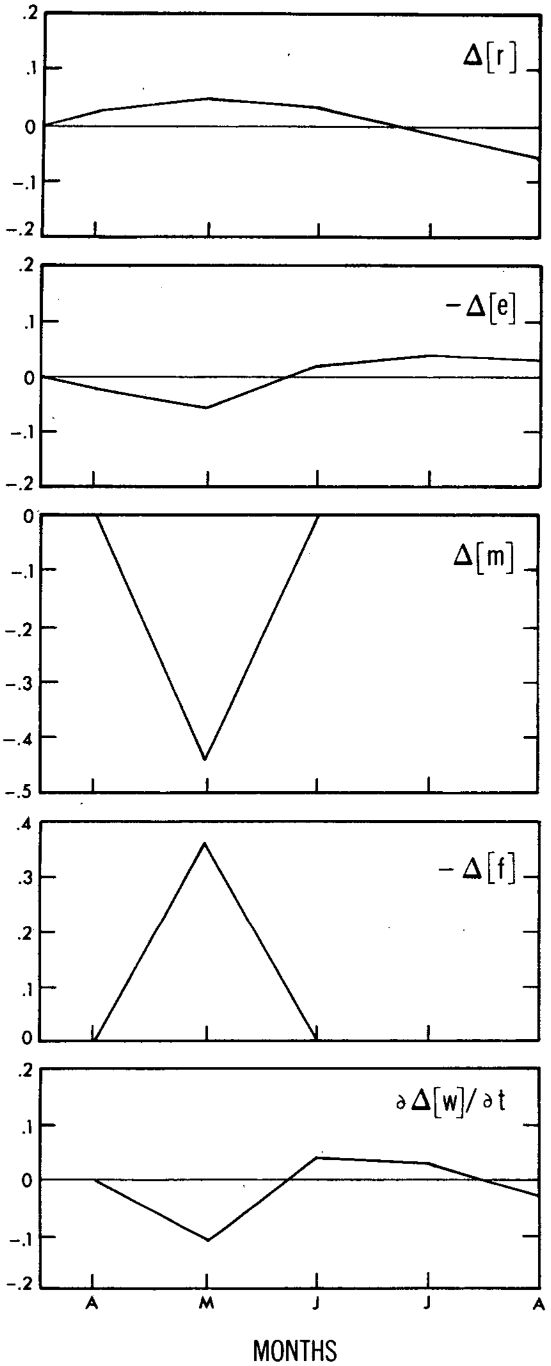

FIG. 11. Time variation of the zonally-averaged monthly mean differences in the water balance components at 75°N latitude over the continent. The differences for March are set equal to zero. Units are in cm day^{-1}.

the two experiments are rather small, the decrease of snowmelt leads to a large reduction of runoff. After June, the difference of snowmelt and runoff between the two experiments becomes zero. This is, again, reasonable because after June there is no snow to melt and the soil is also so dry that there is little or no runoff at this time. Thus, after June the difference in the rate of soil moisture change (i.e., $\partial\Delta[w]/\partial t$) results mainly from the difference of rainfall $\Delta[r]$ and evaporation $\Delta[e]$ between the two experiments.

It was previously shown that the maximum reduction in $\partial[w]/\partial t$ occurred in May. This depletion of soil moisture does not entirely result from the decrease of snowmelt as it may appear. In May the average decrease in the rate of snowmelt reaches 0.44 cm day^{-1}. But the rate of runoff also indicates a large decrease in May amounting to about 0.35 cm day^{-1}. The difference between the two yields a net rate of depletion of soil moisture of about 0.09 cm day^{-1}. This gives a total decrease of soil moisture content in May of approximately 2.8 cm. The rest of the depletion of soil moisture in May is 0.7 cm (3.5 − 2.8 = 0.7 cm) which results from the excess of evaporation over rainfall in the RS experiment. This is about one fourth of the amount of soil moisture depletion which is attributed to the decrease of snowmelt.

We now turn our attention to the curves for $-\Delta[e]$ which shows that evaporation increases in spring from the standard experiment to the RS experiment, but decreases in summer. This phenomenon has already been noticed previously. The curve for $\Delta[r]$ shows an opposite trend as that of $-\Delta[e]$ but the increase of rainfall continues until June after which it decreases. The net contributions from both $\Delta[r]$ and $-\Delta[e]$ are rather small during this entire time. In high latitudes where surface temperature is relatively low, the so-called Bowen's ratio (i.e., the rate of sensible to latent heat flux from the earth's surface) is larger than the corresponding ratio in lower latitudes. This implies that the fraction of surface ventilation due to latent heat flux reduces with increasing latitude provided that soil wetness does not change with latitude. Thus, the evaporation rate is relatively small in high latitudes. Accordingly, the difference in evaporation rate between the two experiments also remains small throughout the course of the numerical experiments. This is an important reason why soil moisture at high latitudes in the RS experiment remains significantly less than the corresponding soil moisture in the standard experiment during the entire summer season. As discussed in Section 7, this dry anomaly is statistically significant at the 90% confidence level for practically the entire period of integration.

6. The thermal response and the change of the zonal circulation

The change of the heat balance near the ground due to removal of snowcover will influence the thermal structure of the atmosphere near the ground. Fig. 12 shows the latitude–month distribution of the zonal

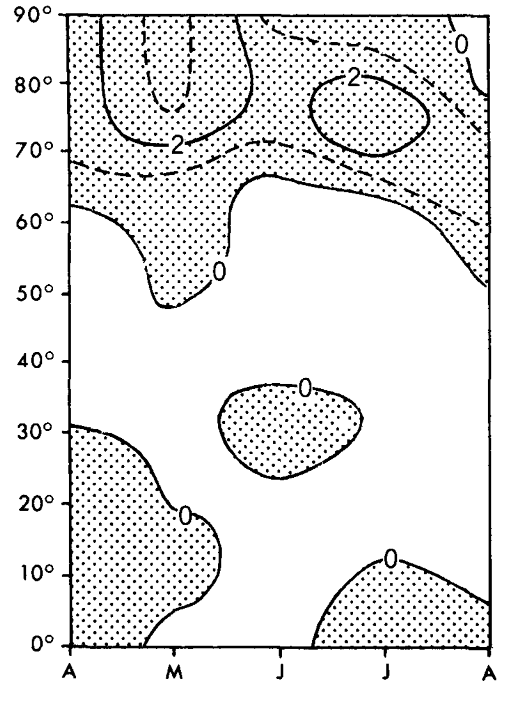

FIG. 12. Latitude–time distribution of zonal mean difference of surface air temperature between the RS and standard experiment over the continent. Units are °C.

mean difference of surface air temperature between the RS and standard experiments. According to this figure the surface air temperature in the RS experiment is higher than that of the standard experiment north of 60°N for the five-month period. The maximum difference appears between April and May reaching a value of almost 4°C in the polar region. However, the increase of surface air temperature resulting from snowcover removal is considerably less than that of the surface ground temperature (see Fig. 8). Therefore, the turbulent flux of sensible heat from the ground to the atmosphere will be increased due to the removal of snow cover (see curve labeled—*SH* of Fig. 6). Further, the comparisons of Fig. 8 and Fig. 12 reveal the similarity of the patterns of the two differences. In other words, the negative and positive areas of the two figures have, for all practical purposes, a one-to-one correspondence. This similarity is not due to the averaging of the eight experiments conducted in this study, since each set of runs shows the same tendency.

In our present study, the systematic increase in surface air temperature is limited to the region of snowcover removal, i.e. high latitudes as may be seen in Fig. 12. This figure also shows that the magnitude of its increase decreases southward from the pole causing a decrease in the zonal mean temperature gradient in the lower atmosphere in middle and high latitudes relative to the model atmosphere for the standard experiment. To illustrate this phenomenon, Fig. 13 is presented which shows the latitude–height distribution of zonal mean temperature difference over the continent between the RS and standard experiments for the month of May. This figure indicates that the magnitude of the temperature increase (i.e. positive difference) decreases smoothly southward from the pole to about 50°N in the lowest three finite difference levels of the model. This figure also shows that, poleward of 50 degrees latitude, the meridional gradient of the zonal mean air temperature decreases upward from the ground surface to ~700 mb.

It is of interest that the pattern of atmospheric temperature change near the model surface in high latitudes is qualitatively similar to the distribution of the CO_2-induced warming obtained by Manabe and Wetherald (1975, 1980). As explained in these two studies, this large warming near the earth's surface is mainly due to the removal of highly reflective snow

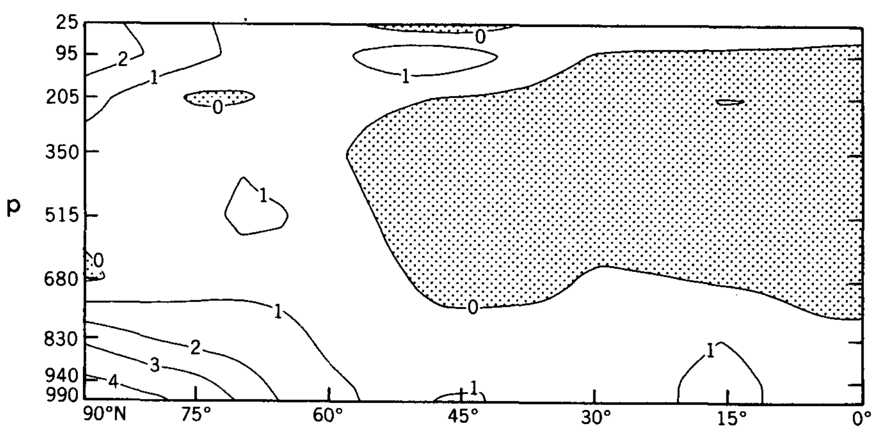

FIG. 13. Latitude–height (pressure, mb) distribution of zonal mean difference of temperature (over land only) for the month of May between the RS and standard experiments. Units are °C.

cover and a stable stratification which limits vertical mixing to the lowest layers of the model atmosphere. Similar processes are operating in the present study.

To respond to this change of thermal structure, the zonal mean westerlies should show a corresponding change. To be consistent with the thermal wind relationship, the zonal mean westerlies must decrease, i.e. there should be a negative difference of zonal mean westerly wind speed between the RS and standard experiments which, in fact, is shown in Fig. 14. According to this figure, there is a decrease of the zonal mean wind of the RS experiment relative to the standard experiment. The maximum decrease of zonal mean wind is more than four meters per second at 65°N latitude and a height of 350 mb. This is not a large decrease in absolute terms. However, since the high latitude westerly current is, itself, not very strong and the distribution of the change of zonal wind (see Fig. 14) has a large scale pattern, this change may have a significant dynamical implication.

7. Statistical significance

To evaluate the statistical significance of the results presented in this study, a Student's "t" test as suggested by Chervin and Schneider (1976) was performed on two key difference distributions. These are the time–latitude difference distributions of surface ground temperature and soil moisture $\Delta[w]$, respectively. The procedure for evaluating the Student's "t" distribution at the 90% confidence level is described in MWS and is applied to the present results in exactly the same way as in that previous work.

Figure 15 shows the Student's "t" analysis as performed on the time–latitude difference distribution of surface ground temperature (Fig. 8). Here, the t-test values are normalized by the critical value of t which corresponds to the 90% confidence level for a sample size of 8 or 14 degrees of freedom. Therefore, any value of this normalized "t" distribution greater than the absolute value of unity represents an area where the null hypothesis that the indicated difference is zero may be rejected at the 90% confidence level.

According to Eq. 15, the positive differences of surface ground temperature in high latitudes shown in Fig. 8 are statistically significant at the 90% (or higher) confidence level. Therefore, removal of snow cover consistently produces warmer surface temperatures in high latitudes.

Similarly, according to Fig. 16, the t-test values show the negative differences of soil moisture at high latitudes (Fig. 9) to be significant at the 90% (or higher) confidence level. Thus, the removal of snow cover consistently has a drying effect for the entire spring and summer seasons. It should be emphasized that the area of statistically significant change is limited to the high latitude zone where the snow cover was removed, with smaller changes elsewhere in the computational domain. This is not surprising in view of the nature of the perturbation (i.e. removal of snowcover) investigated in this study.

8. Conclusions and suggestions for further research

The most significant finding of this study concerning the impact of snow cover on hydrology is that a temporary but complete removal of snow cover can bring about a significant reduction of zonal mean soil moisture for the following spring and summer seasons. It is found that a large negative difference of soil moisture between the RS and standard experiments develops during the month of May and is caused by two factors. These are:

1) Removal of snow cover and, therefore, elimination of moisture available for snowmelt which would otherwise have filled the soil to saturation or else produced runoff.

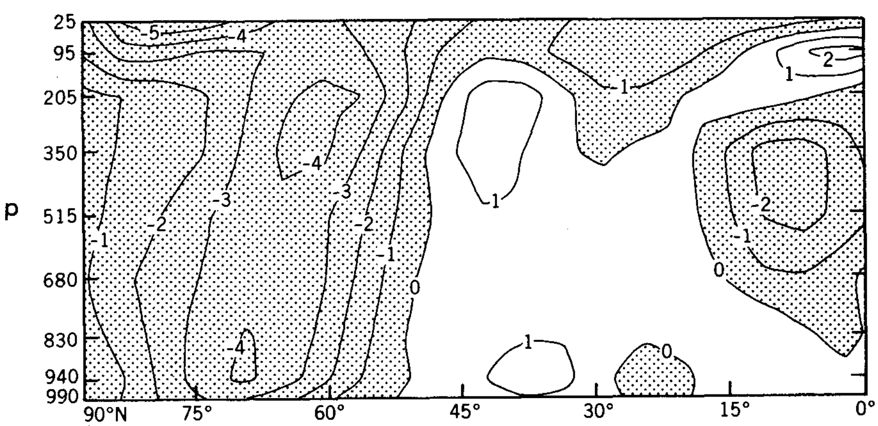

FIG. 14. Latitude–height (pressure, mb) distribution of the zonal mean difference of zonal wind velocity (over land only) between the RS and standard experiments for the month of May. Units are in m s^{-1}.

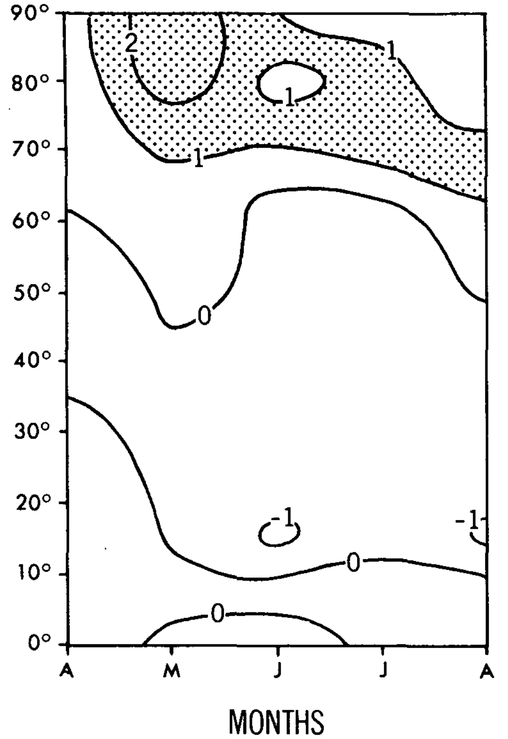

FIG. 15. Student's "t" analysis performed on the difference of surface ground temperature between the RS and standard experiments over the continent. Values are normalized by the critical value of t at the 90% confidence limit for a sample size of eight. Shaded area denotes region where the surface ground temperature difference is statistically significant to the 90% level or greater.

2) Increased evaporation from the soil surface due to the removal of snow cover.

It is further found that the negative difference of soil moisture is present for the entire spring and summer seasons in high latitudes. In this region, the surface temperature is relatively low and, therefore, the Bowen's ratio is larger than that in lower latitudes. This implies a reduced ventilation of the surface by latent heat flux in high latitudes as compared with lower latitudes. Therefore, the difference in the evaporation rate between the two experiments remains small for the entire period of integration. Thus, an initially induced anomaly of soil moisture in high latitudes persists for a long period of time.

As noted in this study, the cause of early summer (June and July) dryness originates in early spring. According to MWS, the increase of CO_2 content may also induce a summer dryness in high latitudes primarily due to an earlier end of the snowmelt season and therefore an earlier beginning of the period of soil moisture reduction caused by enhanced evaporation. Since snow cover (and hence the equivalent snowmelt) is simply removed in early spring in the present investigation, the above result from MWS is relevant to the current analysis.

The above discussion shows that an instantaneous complete removal of snowcover in early spring can produce a hydrological climatic effect which can last as long as four to five months. What we have simulated in this study, although hypothetical, may be relevant to a real situation. It is possible that in a given winter or early spring season, a large area of snowcover in middle and or high latitudes may disappear due to a premature early spring warming. It is also possible that an entire winter and early spring season could occur without any substantial snowfall over a very large middle or high latitudinal region. These situations would be quite similar to our present theoretical study.

In view of the results obtained in the present investigation, it may be worthwhile to perform more experiments. For example, experiments with a large increase (rather than a decrease) in area and depth of snowcover could be performed. Furthermore, since the model used in this current study employs an idealized geography, models with more realistic geography could also be used. One possible experiment with this type of model would be to investigate the consequence of an extreme anomaly of snowcover. Other possibilities include experiments with extensive and deep snowcover or with very little snowcover placed over the Tibetan plateau. The Tibetan plateau

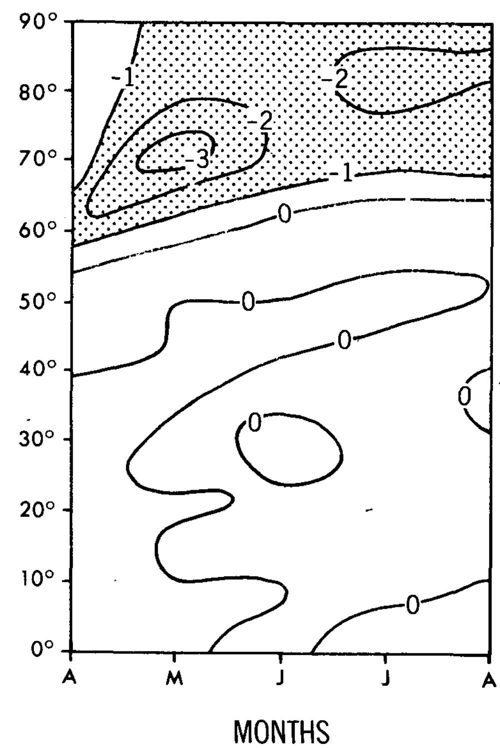

FIG. 16. As in Fig. 15 except for the difference of soil moisture between the RS and standard experiments. Shaded area denotes region where the soil moisture difference is statistically significant to the 90% level or greater.

experiment would be interesting for two reasons: first, because of its relatively low sub-tropical latitude and second, because of its great height above sea level, the average height ~ 5 km.

Although the changes as produced by this study are mainly confined to the high latitude region, it is possible that a similar phenomenon may occur in middle latitudes as well. For example, an earlier snow melt season for a given mid-latitude region accustomed to heavy snowfall could cause enhanced dryness there during the succeeding spring and summer seasons as simulated in this investigation.

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