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Sensitivity experiments with a seasonal energy-balance climate model

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सार — सौर स्थिरांक में परिवर्तन के संदर्भ में स्थलीय जलवायु की संवेदनशीलता का अध्ययन किया गया है। इस अध्ययन में एक ऋतुनिष्ठ, दो स्तरीय ऊर्जा-संतुलन निदर्श का उपयोग किया गया है। संवेदनशीलता के अक्षांशीय तथा ऋतुनिष्ठ वितरण पिछले अन्वेषणों के परिणामों से आमतौर पर मेल खाते हैं। विशेष तौर पर, ऋतुनिष्ठ संवेदनशीलताओं के निर्धारण में समुद्री-हिम-तापीय जड़त्व-पुनर्निवेशन के प्राबल्य की रोबोक (1983) अवधारणा सिद्ध हो जाती है।

ABSTRACT. A study of the sensitivity of the terrestrial climate with respect to change in solar constant is performed using a seasonal, two layer energy-balance model. The latitudinal and seasonal distributions of sensitivity are found to be in general agreement with results of previous investigations. In particular, Robock's (1983) recent hypothesis of the dominance of the sea ice-thermal inertia feedback in determining the seasonal sensitivities is confirmed.

1. Introduction

The effect of change of the solar constant on the terrestrial climate is of considerable interest to the scientific community.

Lately, as a result of the work of Budyko (1969) and Sellers (1969) there has been a surge of interest in the subject. Although these authors used different modelling methods, both of them came to the rather startling conclusion that a reduction of about only 2% in the solar constant is enough to bring about an ice covered earth; *i.e.*, the current climate is dangerously close to a catastrophic deep-freeze. However, their models solved only the surface energy balance equation and the parameterization of the basic physical processes were simplified. Since then several other investigators have used supposedly more realistic climate models to investigate the same problem (Stone 1978; Ohring & Adler 1978; Held 1978; Oerlemans & Vanden Dool 1978); probably the most physically comprehensive of them all being the General Circulation Model (GCM) experiment of Wetherald and Manabe (1975). They used an atmospheric GCM with mean annual forcing, a limited computational domain, no heat transport by ocean currents, and fixed cloudiness. Although their results agreed qualitatively with that of Budyko (1969) and Sellers (1969), they found the climate to be much less sensitive to change in solar constant. Another interesting result of their work was the extreme sensitivity of the hydrologic cycle to change in solar constant. Held (1978) has used a Statistical-Dynamical model (SDM) to perform a detailed analysis of the result of change in solar constant, whereas Peng *et al.* (1982) have used SDM to carry out solar constant experiment and arrived essentially at the same conclusion as Wetherald & Manabe (1975).

2. The model

The model described in details in Birchfield *et al.* (1982) is a time-dependent, seasonal atmosphere-hydrosphere model that solves the energy-balance equations at two atmospheric levels in addition to the surface energy balance equation. Temperature is the prognostic variable for all these equations; therefore the time dependent static stability can be predicted. The model is zonally symmetric and the Crank Nicholson method is used to solve the model equations on a latitudinal grid size of 4 deg. with 60 time steps per year. Each grid area is appropriately divided into land and occean parts using present day distribution. The 'ocean' is a 120 metre deep isothermal mixed layer; meridional heat flux in the ocean is not taken into account. The only heat transport in the meridional direction is accomplished by the atmospheric heat flux and this is simulated by a linear diffusion process. Since the model does not have a deep ocean component, the relaxation time well below 100 years. Typically it takes between 25 and 50 years for the seasonal cycle to reach equilibrium.

There are two ways in which energy transfer can take place in the model between the two atmospheric levels and the surface. One of them is the radiative transfer. Both short-wave and long-wave fluxes depend on radiation calculations performed once and for all, and these fluxes are obtained in the model as tabulated functions of model temperatures and solar zenith angles. The other way of energy transfer in the vertical direction in the model is moist convective adjustment; whenever the lapse rate falls below a local critical level, energy exchange takes place between layers so that the lapse rate is equal to the moist adiabatic lapse rate, without changing the mean temperature of the two layers.

At each latitude, surface energy balance computations are performed separately for land and ocean, and the surface flux to the atmosphere is the weighted sum of the separate contributions. When the temperature of the mixed layer falls below 273 deg. K, sea-ice forms with freezing from below and a sea-ice budget is invoked. The model does not have a closed hydrologic cycle; it is assumed that all the water that evaporates at a given latitude also precipitates at the same latitude. Evaporation to the atmosphere is governed by the vertical gradient of the water vapour between the surface assumed saturated and the atmospheric boundary layer assumed 80% saturated. Snow falls whenever the surface air temperature falls below 273 deg. K and a snow budget calculation is performed at the same time. The rate of snowfall is determined by latitude only.

The model has two other distinctive features. First, atmospheric diffusion is assumed to take place on isobaric surface. Second, due to the separation of land and ocean fractions at each latitude, an assumption had to be made relating the potential temperature profiles over land and ocean so that they could be combined; it was assumed that the potential temperature over a sigma surface was same over land and ocean.

The 'control' value of the solar constant used in these experiments was 1367 watt/metre^2.

3. Results

The model has two basic versions: with and without ice albedo feedback. In the version with ice albedo feedback the model computes the surface albedo at each grid point depending on the presence or absence of snow and sea-ice, separately for land and ocean. Bare land has an albedo of 0.14; if the land is covered with snow then its abledo is set to the old snow albedo (.60) or the new snow albedo (.85). The albedo of the bare ocean water is .07; this is one of the shortcomings of the model, because the ocean albedo is known to have a strong dependence on zenith angle. When sea-ice is present, the ocean albedo is either that of bare sea-ice (.60) or, in case of snow cover on sea-ice, equal to that of old or new snow. Snow falls when the surface air temperature is less then 273 deg. K. If the mixed layer temperature is less than 273 deg. K, sea-ice is assumed to cover the ocean.

In the version without the ice-albedo feedback the albedo of land is .14 and that of the ocean is .07 for all surface conditions.

Time dependent computations were carried out for the version with ice-albedo feedback till equilibria were reached, for the following values of solar constant perturbation: 0% (control), + 2%, —2%, +4%, +6%. For this version, a perturbation of —4% drives the earth's surface to an ice-covered state. For the sake of comparison, the version without ice-albedo feedback was run for the following perturbations: 0% (control),+ 4% and —4%.

Interestingly, the time to reach equilibria were found to depend on the value of the solar constant. This result is in agreement with that of Held & Suarez (1974) about the stability of the equilibrium states. They showed theoretically that as the solar constant is reduced toward the critical value for large ice cap instability, the relaxation time increases. According to their analysis the relaxation time is related both to the latitude of the ice cap boundary and the magnitude of the solar constant at that boundary. This result is also in agreement with the results of the GCM experiments of Wetherald and Manabe (1975) even though their model is physically much more comprehensive.

The times taken by the model to reach equilibrium are given in Table 1. It is seen that these times (usually between 25 and 50 years) are determined to a large extent by the solar constant, as explained before. These values are probably somewhat high because of the

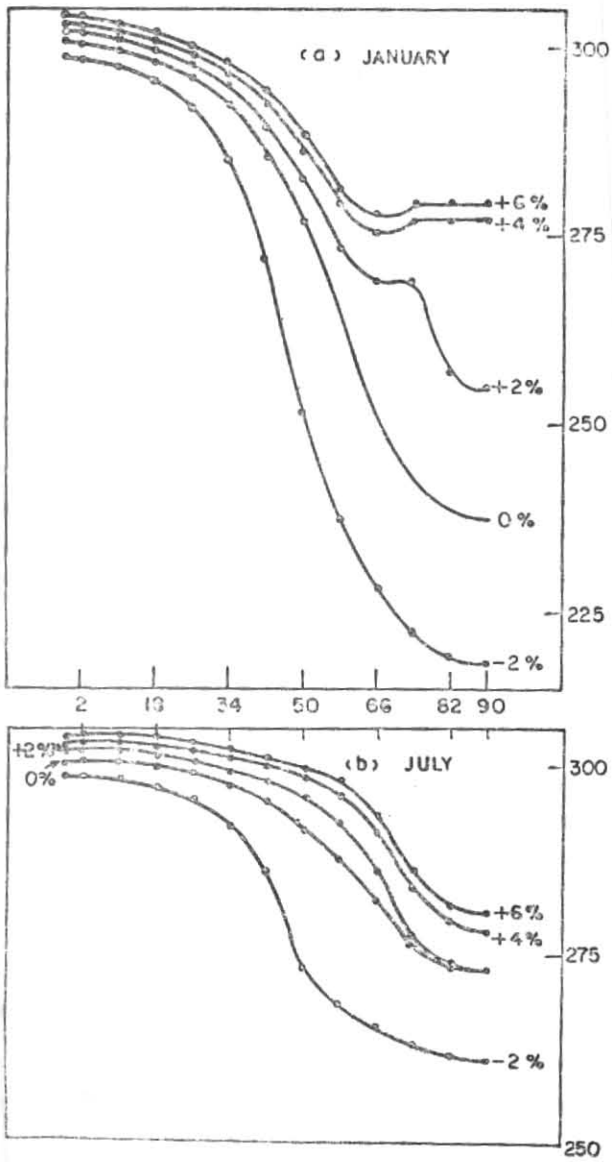
TABLE 1

Time taken to reach equilibrium in time step units

Change in solar constant	January	July
Control (0%)	2090	2060
+2%	2090	2060
−2%	3050	3020
+4%	1970	1940
+6%	1550	1520

large value of the mixed-layer depth (120m) used in the model. Ice (like land) does not have any heat capacity in these atmosphere-hydrosphere equilibrium computations.

The equilibrium latitudinal distributions of surface temperatures for various valuts of silar constant are shown in Figs. 1(a) & 1(b). Hemispheric mean surface temperature and surface air temperature are listed in Table 2. Several interesting features are revealed by these diagrams. First, the sensitivity of the surface temperature of change in solar constant is much higher at higher latitudes than that at the low latitudes. This is directly illustrated in Figs. 2(a) & 2(b). For comparison, the sensitivities for the model version without ice-albedo feedback are shown on the same diagrams (marked with crosses) for the case of + 4% increase in solar constant. The large values of sensitivity at the high latitudes can be explained by the presence of snow and ice at those latitudes. This is the result of ice-ablbedo feedback : as the surface temperature decreases, the area covered by snow and ice increases leading to further cooling of the surface because the albedoes of snow and ice are much higher than that of bare land. Figs. 2(a) and 2(b) show that the ice albedo feedback increases the surface temperature sensitivity by more than an order of magnitude in January, and somewhat less in July. In addition, the large stability of the atmosphere at high latitudes amplifies the surface temperature sensitivity by suppressing vertical mixing. This is in agreement with the GCM results of Wetherald & Manabe (1975) and also with the nine layer SDM experiments of Peng *et al.* (1982). They noted that the atmospheric temperature response increases with height at low latitudes due to strong cumulus convection, and decreases with height at high latitudes due to the above mentioned stability. In our model, the low latitude convection has been represented by the moist adiabatic adjustment process.

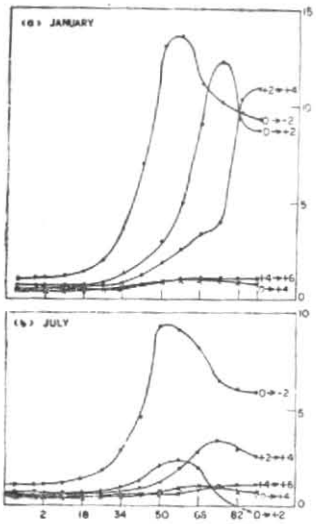


Figs. 1(a & b). Surface temperature for (a) January and (b) July (in deg. K) plotted against latitude (degrees) for the Northern Hemisphere, for various values of the solar constant. 0% denotes control value (1367 W/M²); others are the changes on the control value

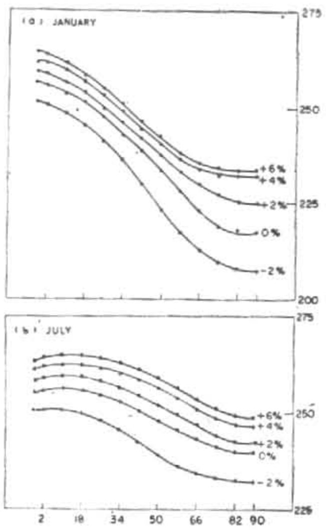
TABLE 2

Mean surface temperature and surface air temperature for the Northern Hemisphere (deg. K)

Change in solar constant	*T* (Air)	*T* (Surface)
Control (0%)	289.1786	290.2638
+2%	292.8136	293.9945
−2%	280.8242	281.4586
+4%	295.6851	296.8782
+6%	296.9595	298.0745



Figs. 2(a&b). Sensitivity of surface temperature (in deg. K) for : (a) January & (b) July plotted against Northern Hemisphere latitude (degrees), for transititions between various values of solar constant. Ordinate denotes normalized sensitivity, *i. e.*, sensitivity per unit percentage change in the solar constant. The special case of no ice-albedo feedback is denoted by crosses



Figs. 3(a&b). Same as in : (a) Fig. 1 (a) & (b) Fig. 1 (b) but for column mean temperature

TABLE 3

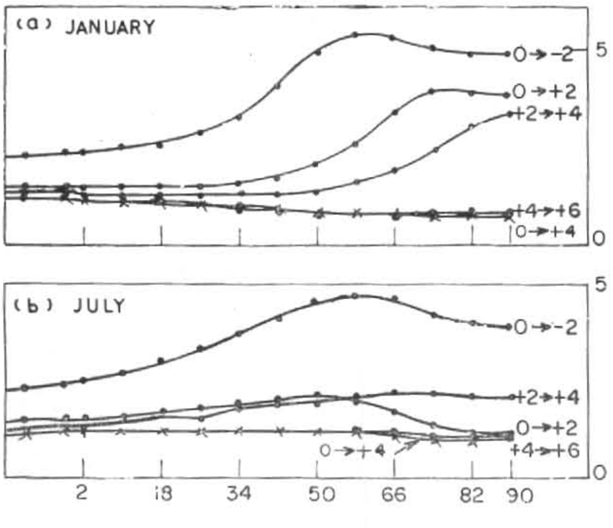
Mean change in surface temperature and surface air temperature for Northern Hemisphere (deg. K)

Transitions	T (Air)	T (Surface)
$0 \rightarrow +2$	3.635	3.731
$0 \rightarrow -2$	8.354	8.810
$+2 \rightarrow +4$	2.872	2.884
$+4 \rightarrow +6$	1.274	1.196

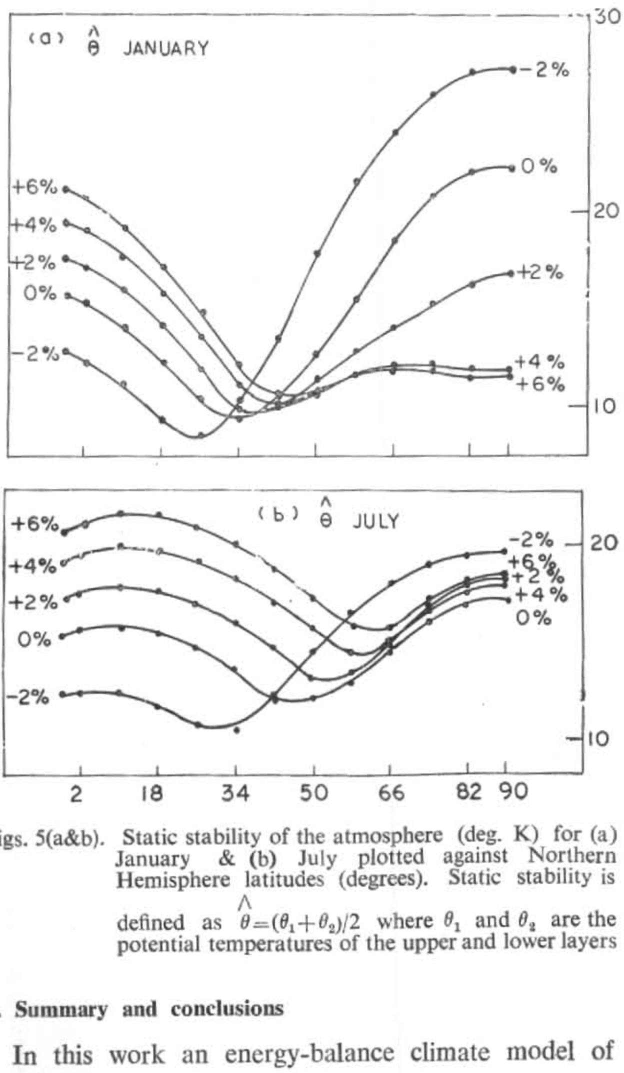
The second interesting feature is the increase in sensitivity with decrease in the value of the solar constant. Such a non-linear relationship has already been noted by Wetherald & Manabe (1975), Peng *et al.* (1982). Held and Suarez (1974) have theoretically argued that this relationship depends on the details of the parameterization of physical processes in the model. This probably explains the hemispheric mean sensitivities shown in Table 3. Although the sensitivity for 2% increase in solar constant is not very different from sensitivity obtained for the same case by others (Wetherald & Manabe 1975, Table 1; Peng *et al.* 1982, Table 3). the sensitivity for 2% decrease is about a factor of two larger for the present model. This situation probably indicates explosive growth of ice and snow cover, although the solar constant has to be still lowered for arriving at the 'white earth' solution. Budyko (1969) and Sellers (1969) obtained similar very high sensitivities, although later investigators (North 1975; Lian & Cess 1977; Coakley (1979) have found lower values.

Another interesting feature is the increased sensitivity during winter compared to summer in the polar regions. This agrees with the result of Robock (1983) who used a seasonal energy-balance model to test a new parameterization of snow and ice area and albedo based on recent satellite data. According to him, the seasonal sensitivity pattern is largely determined by the sea-ice thermal inertia feedback. When the ice area changes due to change of the surface temperature, the thermal inertia of the ocean-ice-atmospher system changes. This feedback dominates the albedo feedback (*i.e.,* the snow/ice-area feedback and the snow/ice meltwater feedback) in producing the seasonal contrast.

Mean temperatures of the atmospheric column are shown in Figs. 3(a) & 3(b). The associated sensitivities are shown in Figs. 4(a) & 4(b). It is seen that the seasonal pattern of the distribution is same as that of the surface temperature, although not as variable. The features of sensitivity distribution are qualitatively same as before, *i.e.,* enhanced sensitivity at high latitudes, sensitivity increasing with reduction in solar constant, and higher sensitivity in winter than in summer. However, these are much weaker than before. An explanation of this behaviour is given by Held *et al.* (1981). They noted that the reason that mean air temperatures at high and low latitudes are well coupled w.r.t. solar constant change whereas the surface temperatures are reasonably insulated is that there is a profound change in the distribution of static stability. This is confirmed by Figs. 5(a) & 5(b) which show the latitudinal distribution of stability. As discussed by Held *et al.* (1981), the boundary between regions where stability decreases with decrease in temperature (due to decrease in stability of moist adiabat) and increases with decrease in temperature moves equatorward as the temperature decreases, because the zone of moist convection shrinks. This also explains the contrast in seasonal structure of the stability distribution: the minima in stability move farther from equator in summer because of the enhanced region of moist convection. The fact that there is a quite strong minimum in the stability at mid latitudes denotes a weakness of the present model; vertical energy transfer from lower to upper layer is too weak at those latitudes. However, there does not seem to be any obvious way to correct this problem within the framework of an energy-balance model.



Figs. 4(a&b). Same as in : (a) Fig. 2(a) & (b) Fig. 2(b) but for column mean temperature



Figs. 5(a&b). Static stability of the atmosphere (deg. K) for (a) January & (b) July plotted against Northern Hemisphere latitudes (degrees). Static stability is defined as $\hat{\theta}=(\theta_1+\theta_2)/2$ where θ_1 and θ_2 are the potential temperatures of the upper and lower layers

4. Summary and conclusions

In this work an energy-balance climate model of intermediate sophistication is used to study the effect of change in solar constant on the terrestrial climate. The key features of the model are the seasonal resolution and the separation of atmospheric and surface energy-balance computations, so that the static stability can be predicted. The results are in general agreement with those of other investigations, *i.e.,* that sensitivity of the surface temperature with respect to change in solar constant increases towards the pole, increases with decrease in solar constant, and is higher in winter than in summer. However, the threshold for the large ice cap instability is very low; in fact it is close to the values obtained by Budyko (1969) and Sellers (1969). This probably is due to the particular form of the ice-albedo parameterization used. However, there are many other simplifications which would probably affect the model's performance, *e.g.,* uniform cloudiness, no oceanic heat transport, absence of a closed hydrologic cycle, constant diffusivity, to name a few. In this context, it should be mentioned that present of high-latitude topograph and mid-latitude topograph increases the same model's sensitivity considerably, as demonstrated by Birchfield *et al.* (1982) and Birchfield and Weertman (1983). Inclusion of snow/ice meltwater feedback also increases the sensitivity considerably (Birchfield and Weertman 1982).

Although less physically comprehensive than general circulation models, intermediate models like the present one will probably continue to play important role in climate research because they are less expensive, more tractable and easier to analyse.

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