Effects of Nontropical Forest Cover on Climate

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ABSTRACT

The albedo of a forest with snow on the ground is much less than that of snow-covered low vegetation such as tundra. As a result, simulation of the Northern Hemisphere climate, when fully forested south of a suitably chosen taiga/tundra boundary (ecocline), produces a hemispheric surface air temperature 1.9 K higher than that of an earth devoid of trees. Using variations of the solar constant to force climate changes in the GLAS Multi-Layer Energy Balance Model, the role of snow-albedo feedback in increasing the climate sensitivity to external perturbations is reexamined. The effect of snow-albedo feedback is found to be significantly reduced when a low albedo is used for snow over taiga, south of the fixed latitude of the ecocline. If the ecocline shifts to maintain equilibrium with the new climate—which is presumed to occur in a prolonged perturbation when time is sufficient for trees to grow or die and fall—the feedback is stronger than for a fixed ecocline, especially at high latitudes. However, this snow/vegetation–albedo feedback is still essentially weaker than the snow-albedo feedback in the forest-free case.

The loss of forest to agriculture and other land-use would put the present climate further away from that associated with the fully forested earth south of the ecocline and closer to the forest-free case. Thus, the decrease in nontropical forest cover since prehistoric times has probably affected the climate by reducing the temperatures and by increasing the sensitivity to perturbations, with both effects more pronounced at high latitudes.

1. Introduction

Strong concern has been expressed in recent years that human activities may change our climate, with significant consequences for the agriculture of some regions and their habitability. The impact on the land surface has recently attracted increased attention (Sagan *et al.,* 1979). This study concerns one aspect of this problem area, *viz.,* the climatic effects of deforestation of temperate and boreal forests.

Trees significantly reduce the terrain albedo when snow covers the ground. Measurements (Federer, 1968, 1971) indicate the magnitude of this effect. Modeling (Federer, 1971; Otterman, 1981, 1984) helps us to interpret the measurements and to assess the influence of various parameters on the albedo. The climatic significance of temperate or boreal forests, resulting from their low winter albedo, is analyzed by simulating two extreme situations, one in which the earth is fully forested south of a taiga/tundra boundary and one in which the earth is devoid of trees. For these two situations, we examine the annually averaged zonal surface air and land surface temperatures. In addition, we compare the sensitivities of these two climates to perturbations in the solar constant. These sensitivities differ because the strength of the snow–albedo feedback differs in the two situations.

When climatic cooling occurs in response to an external or internal perturbation, the snow–ice cover extends over larger areas and is of longer seasonal duration. It increases the planetary albedo and reduces the absorption of solar radiation by the earth, thus enhancing the cooling beyond its primary effect. This mechanism, known as the snow/ice–albedo feedback, increases the sensitivity of the climate to perturbations. It has been extensively studied in a hierarchy of climate models, including simple energy balance models (Budyko, 1969; North *et al.,* 1981), a general circulation model (Wetherald and Manabe, 1975) and more recently, the GLAS Multilayer Energy Balance Model (MLEBM) (Peng *et al.,* 1982). In this last study, the assessments of the snow/ice–albedo feedback by various workers are compared.

We reexamine the sensitivity of climate to the solar constant with a version of the MLEBM, in which land and ocean temperatures in each latitudinal zone are separately computed. We first simulate the climate in which the albedo over a snow-covered terrain is computed as a function only of the surface temperature, without regard to vegetation height. We next take into account the albedo of a forested surface with snow on the ground being much lower than that of snow-covered tundra, and then study the sensitivity of climate to changes in the solar constant (as an example of a forced perturbation) under two alternative situations: 1) a taiga/tundra boundary (ecocline) fixed at latitude with −5°C yearly temperature under the present solar

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constant, and 2) an ecocline at a constant climatic, but variable geographic, location. The former case would be appropriate for short-term climatic perturbations, while the latter case would apply to perturbations that last at least as long as the time it takes for trees to develop or die and topple.

2. The modified multilayer energy balance climate model

The basic climate model is fully described by Peng *et al.* (1982) and only a brief overview is provided here. The Northern Hemisphere is divided into nine latitudinal zones, each 10° wide, and nine layers in the vertical: eight tropospheric layers with equal pressure intervals and one stratospheric layer above the 200 mb level. The atmospheric temperatures are computed from the thermodynamic energy balance equation, which includes the heating due to solar and IR radiation, latent heat release, the mean meridional circulation in the tropics, and large-scale transient eddies in the extratropical regions. Mesoscale and small-scale eddies contribute to the heating only in the vertical direction. Different from the original version of the model, the surface and the boundary layer temperatures are computed separately over land and oceans based upon the heat fluxes at the surface and in the oceans.

The model is annually averaged with no seasonal cycle. Thus, all the reported temperatures represent annual means. The relative humidity is specified according to climatological data. Clouds are classified into five types with amounts, heights and optical thicknesses specified.

The annual mean surface albedo R_g in a latitude zone ϕ is computed from

$$R_g(\phi) = \sum_{j=1}^{4} f_j(\phi) \sum_{i=1}^{4} n_{ij}(\phi) R_{ij}(\phi), \qquad (1)$$

where R_{ij} is the albedo of surface type i during season j, n the fraction of the latitude belt covered by each surface type, and f the weighting function that accounts for the seasonal change in the solar radiation incident at the surface. Four surface types are considered: open ocean, land, snow and sea ice. Land albedos are taken from Sellers (1973) and vary with latitude only. Ocean albedos are obtained from Budyko (1974) and vary with latitude and season. The values of f are also obtained from Budyko (1974). At 55°N, for example, they are 0.07, 0.36, 0.42 and 0.15, respectively, for winter, spring, summer and autumn.

In order to account for changes in the albedo of snow and sea ice due to partial melting and accumulation, the albedo is taken to vary with the coverage. A linear dependence is assumed such that the mean albedo of snow varies from 0.5 to 0.7 and the albedo of sea ice varies from 0.4 to 0.6, respectively, as the fractional coverage of snow over land and that of ice over oceans varies from 0 to 1. The fractions of snow and ice cover are computed from the empirical relationships between seasonal snow/ice cover and the annually and zonally averaged surface temperature derived by Chou *et al.* (1981).

3. Conventional high winter albedo (HA) representation

The model is applied to simulate the climates with the solar constant at its present value (denoted by $1.00S_0$) and ±2% changes from this value. The albedo of snow is in the 0.5–0.7 range, depending on the snow cover (see Peng *et al.*, 1982). For $1.00S_0$, the hemispherically averaged surface air temperature is 286.67 K and the land surface temperature at 65°N is 264.13 K. The surface air and land temperatures in latitudinal zones are presented in column 2 of Tables 1 and 2, respectively. The land and the zonal albedos are presented in Table 1.

Simulations of the climate response to +2% and −2% changes in the solar constant are as follows: +3.47 K and −4.32 K in the hemispherically averaged surface air temperatures and +5.83 K and −7.01 K in the land surface temperatures at 65°N. These results are similar to those reported by Peng *et al.* (1982). The responses of land surface temperatures are tabulated in columns 3 and 4 of Table 2 and are plotted in Fig. 1 (curves a).

We refer to our simulation and results in this section as the conventional, high winter albedo (HA) representation, which is applicable to the case of an earth devoid of trees. In the next section, the albedo of forested areas with snow on the ground is assumed to be much lower.

4. Low winter albedo representation (LA) over forests, geographically fixed ecocline

In the above simulation, the albedo of terrain with snow on the ground is in the 0.5–0.7 range, depending on only the zonal temperature. These values are quite inappropriate for the vast regions of the boreal forest (taiga) and the temperate forest, where snow rarely covers the treetops. Snow only infrequently remains on the branches. The conifers of taiga have short springy branches which help to shed the snow. Their overall conical shape also tends to prevent an accumulation (Money, 1972).

Measurements clearly establish the point that forested areas with snow on the ground have much lower albedo than the snow-covered fields. Thus, with a snow cover consisting of 15 cm of two-day-old powder snow on a preexisting 35 cm snowpack, the albedo of hardwood and pine forest is in the range 0.14–0.25, whereas the nearby snow-covered fields have an albedo of 0.72 (Federer, 1968). These tower measurements over the New Hampshire sites are at solar zenith angles of 62–74°. In another study, Federer (1971) reports that in

TABLE 1. Zonal surface air temperatures and albedos for the high winter albedo representation (HA) and the low winter albedo representation (LA) with fixed ecocline for the $1.00S_0$ case. The observed surface air temperatures are from Oort and Rasmusson (1971).

Latitude (°N)	Surface air temperature (K)			Zonal surface albedo		Albedo over land	
	HA	LA	Observation	HA	LA	HA	LA
5	297.76	298.86	299.4	0.064	0.064	0.08	0.08
15	298.04	299.13	298.6	0.080	0.080	0.13	0.13
25	294.68	295.79	295.3	0.109	0.109	0.18	0.18
35	288.73	290.33	289.6	0.118	0.112	0.176	0.162
45	281.84	284.17	281.8	0.171	0.127	0.250	0.165
55	273.98	277.23	275.9	0.272	0.149	0.346	0.247
65	264.52	269.17	268.8	0.494	0.300	0.527	0.301
75	256.28	259.25	261.1	0.606	0.575	0.662	0.624
85	251.66	254.13	253.8	0.606	0.606	0.70	0.70
Mean	286.67	288.58	288.0				

a leafless deciduous forest with snow on the ground, roughly 65% of the incident radiation is absorbed by stems and branches, 15% is absorbed by the snow and only 20% is reflected. Based on these data, as well as recent aircraft measurements in New York State (Kukla and Robinson, personal communication, 1983), the value of the albedo of a forest with snow on the ground is assumed to be 0.25. Such a reduction in the albedo by the dark trees protruding above the bright snow is entirely consistent with the modeling recently developed for a plane with dense protruding plants (Otterman, 1984), which is an extension of the previously developed model for a plane with sparse protrusions (Otterman, 1981).

An albedo of 0.25 is therefore used for all snow-covered land areas south of the taiga/tundra ecocline. This boundary is drawn at the yearly averaged surface air temperature of −5°C based on the data of Whittaker (1975). Its geographic location is at 66°N when simulating a climate with today's solar constant and with the low-albedo (LA) representation south of the ecocline. This simulated ecocline is probably somewhat to the north of the present average location. (The LA simulation is not meant to represent the climate today. Both the present and the prehistoric climates are compared to LA in Section 6.) The land and the zonally averaged albedos are presented in Table 1. Compared to the HA representation, the surface albedos are much lower in the region immediately south of the ecocline.

The LA simulations produce a significantly warmer climate. The hemispherically averaged surface air temperature is now 288.58 K and the land surface temperature at 65°N is 269.35 K, which differ by +1.91 and +5.22 K, respectively, from the HA simulations. The LA results correspond more closely to the present climate, for which the hemispherically averaged surface air temperature is 288.0 K. Surface air temperatures from the $1.00S_0$ LA simulation and the observed data are presented in columns 3 and 4, respectively, of Table 1. At 65°N, the LA surface air temperature is 0.4 K higher than the observed value, whereas the HA temperature (column 2) is 4.3 K lower. The land surface temperatures are presented in column 5 of Table 2.

With the low albedo of snow south of the taiga/tundra ecocline and maintaining the ecocline at a fixed latitude, we simulate the climate with ±2% changes in the solar constant. The resulting changes in land surface temperatures are presented in columns 6 and 7 of Table 2 and are plotted in Fig. 1 (curves b). The changes are now significantly smaller than that in the

TABLE 2. Model surface temperature (denoted by $1.00S_0$) and the responses to ±2% S_0 changes in the solar constant for two different surface albedo representations. The unit of temperature is K.

Latitude (°N)	High winter albedo representation (HA)			Low winter albedo representation (LA): Ecocline fixed at 66°N			LA: Shifting ecocline	
	$1.00S_0$	+2% S_0	−2% S_0	$1.00S_0$	+2% S_0	−2% S_0	+2% S_0	−2% S_0
	Hemispherically averaged surface air temperatures and responses							
	286.67	+3.47	−4.32	288.58	+2.77	−2.67	+3.05	−3.73
	Land temperatures and responses							
5	304.51	+3.10	−3.71	305.63	+2.74	−2.58	+2.91	−3.22
15	303.22	+3.07	−3.74	304.32	+2.74	−2.58	+2.90	−3.20
25	298.59	+3.03	−3.77	299.70	+2.71	−2.58	+2.91	−3.22
35	291.46	+3.33	−4.15	293.16	+2.66	−2.59	+2.84	−3.24
45	283.42	+3.87	−4.97	286.12	+2.67	−2.73	+2.90	−3.65
55	274.54	+4.66	−6.52	278.57	+3.05	−3.23	+3.45	−4.94
65	264.13	+5.83	−7.01	269.35	+3.88	−3.96	+4.74	−8.01
75	255.36	+4.43	−5.19	258.14	+3.63	−3.22	+4.42	−4.98
85	250.67	+4.11	−4.72	253.19	+3.16	−3.21	+3.56	−4.71

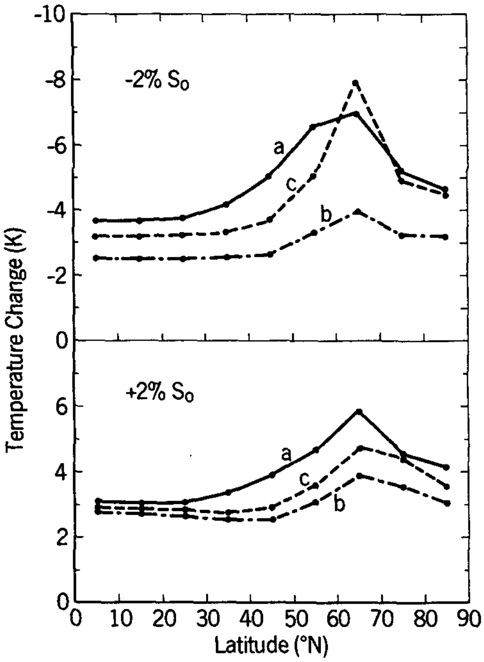

FIG. 1. The latitudinal distributions of the response of the land surface temperature to ±2% changes in the solar constant for (a) high winter albedo representation (HA), (b) low winter albedo representation (LA) with geographically fixed ecocline, and (c) LA with shifting ecocline.

HA simulations: +2.77 K (−2.67 K) in the hemispherically averaged surface air temperature for a +2% (−2%) change in the solar constant, which constitutes a 20% (38%) reduction in the sensitivity, and +3.88 K (−3.96 K) in the land surface temperatures at 65°N, which represents a reduction in the sensitivity by 33% (43%).

Contrary to the results of the HA simulations, the surface temperature is more sensitive to +2% S_0 than to −2% S_0. Since the contrast in surface albedo of snow-covered to snow-free terrain is much larger north of the ecocline (~0.60/0.16) than south of the ecocline (0.25/0.16), the absorption of solar radiation is more sensitive to the snow cover north of the ecocline. For the LA simulations with the fixed ecocline, the change in the snow cover is mostly in the region north of the ecocline for the +2% S_0 case and south of the ecocline for the −2% S_0 case. The climate is therefore more sensitive to the +2% S_0 change than to the −2% S_0 change.

5. Low winter albedo representation over forests, shifting ecocline

In Section 4, the taiga/tundra ecocline is fixed at 66°N. If a climatic perturbation is of a short duration, a year or two, the ecocline can be assumed to persist in the same geographical location. However, if a perturbation persists for a longer period, a few decades or more, the ecocline gradually adjusts its geographic location to the new climate. In the case of a warming, the tree species of the taiga manage to grow where under the cooler climate only the dwarf-shrubs, grasses, mosses and lichens of the tundra grow. In the case of a cooling, the trees of taiga die when exposed to temperatures lower than the acceptable range for their habitat; the roots decay and wind topples the trees. The arctic plants take over, that is, the tundra expands southward.

We now simulate the climate sensitivity to the solar constant by placing the taiga/tundra ecocline at the latitude where the average surface temperature is −5°C. The results are presented in Table 2, columns 8 and 9. The responses of the hemispherically averaged surface air temperatures to +2% S_0 and −2% S_0 are, respectively, +3.05 and −3.73 K. These responses represent significant enhancement in the sensitivity, by factors of 1.1 and 1.4, respectively, when compared to the fixed ecocline simulations. The responses in the land surface temperatures at 65°N, where the taiga/tundra ecocline shifts north and south, are now +4.74 and −8.01 K. The enhancement factors are 1.2 and 2.0 when compared to the fixed ecocline simulations. The much larger effect of the shifting ecocline in the case of the cooling can be simply explained by the extensive land areas south of the ecocline from which the taiga retreats. On the other hand, in the case of the warming, the areas for the expansion of the taiga are limited by the Arctic Ocean. The latitudinal distributions of the responses are plotted in Fig. 1 (curves c). The enhancement factors of these responses as compared to the simulation with a fixed ecocline are plotted in Fig. 2.

It should be noted in Fig. 1 that for the +2% S_0 case, the climate is much more sensitive for HA than for LA whether with a fixed or with a shifting ecocline. Thus, the forests endow our climate system with a

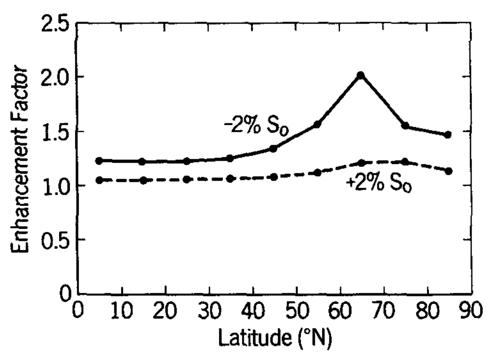

FIG. 2. Enhancement of the response of land temperature to ±2% changes in the solar constant when the tundra/taiga boundary (ecocline) is allowed to shift with climate. As an example, the response of land temperature at 65°N to a −2% change in the solar constant is −3.96 K for the case with ecocline fixed at 66°N; it is −8.01 K with the ecocline shifting with the latitude of −5°C. The enhancement is thus a factor greater than 2.

strongly enhanced stability to a warming perturbation. The situation is more complicated when considering a cooling perturbation, $-2\%\ S_0$. The conventional HA produces higher sensitivity than LA at all latitudes when the ecocline is fixed, but only up to (and including) 55°N when the ecocline shifts with climate. At 65°N, with the shifting ecocline, the response is -8.01 K which is 1 K higher than the result of the conventional HA representation. Thus, the stabilizing effect of forests is interrupted at the region where tundra expands.

6. Summary and implications for changes in the earth's climate

The effect of snow on the surface albedo is much smaller in a forest than over a region of low vegetation. When one takes into account the existence of trees in a fully forested earth south of a taiga/tundra boundary (ecocline), the hemispherically averaged surface air temperature increases by nearly 2 K and the land surface temperature at 65°N by more than 5 K, compared to an earth devoid of trees.

In addition to their effect on mean climate conditions, boreal and temperate forests have a stabilizing effect on the earth's climate by reducing the snow–albedo feedback. Compared to an earth devoid of trees, the mean hemispheric surface temperature of an earth fully forested south of the ecocline would be 20% less sensitive to a short-term warming perturbation ($+2\%\ S_0$) and 38% less sensitive to a short-term cooling perturbation ($-2\%\ S_0$). In a long-term perturbation, where the ecocline is allowed to shift to a new equilibrium, the corresponding reduction in sensitivity is 12% and 14%, respectively.

The long-term climatic perturbations on a time scale of a few decades entail a vegetation–albedo feedback. The tree line, which we refer to as the taiga/tundra ecocline, shifts with some delay in a direction defined by the new climate. The shift continues until the climate and the ecosystems reach a new equilibrium. The tree line is then at a new geographical location, but at −5°C. In the case of warming, the vegetation–albedo feedback enhances the sensitivity of land temperature by a factor ranging from 1.06 in the tropics to 1.2 at 65 and 75°N, as shown in Fig. 2. In the case of cooling, the maximum enhancement factor is 2.02 at 65°N, where the sensitivity is higher than that in the HA case of an earth devoid of forests. It appears, therefore, that at the high northern latitudes, 55–75°N, the presence of forests has little or no stabilizing influence on temperature changes resulting from *a long-term decrease* in the solar constant.

The high sensitivity of the northern latitudes to prolonged cooling perturbations has strong implications for agriculture in these regions. The agricultural season at these latitudes is short. The yields and even the practicability of agriculture are very much climate-dependent (Lamb, 1965). The often-quoted story of the Viking settlement in Greenland at a latitude above 60°N, which apparently failed due to the vagaries of climate, should be mentioned once again. It should be noted, however, that the highest sensitivity is limited in its latitudinal extent at the vicinity of the retreating taiga.

A reduction in the extent of temperate or boreal forest represents a climatic shift closer to HA in the LA to HA interval. The prehistoric conditions of the post-glacial optimum (5000–7000 BP), before much of the forested land was converted to agricultural fields and other land use, were closer to the LA situation than the present situation is. During this prehistoric period, the climate was warmer than today and apparently more stable (Lamb, 1965). Our two representations, the fully forested earth (south of the tundra) and the earth devoid of trees, constitute two extremes that bracket the prehistoric climate and the present climate. A most pertinent question concerning the role man played in the climatic changes from the prehistoric past so far is not answered in our work. To address this question, we need data on the extent of deforestation from prehistoric times.

There are several studies, some quite recent, assessing the extent of the deforestation. Matthews (1983) estimates the deforestation in the temperate forest due to conversion to agriculture as 6.5×10^6 km^2 or 19.5% of its original prehistoric extent. Darby (1956) estimates a much larger reduction for Europe, assessing that about 60% of central Europe was converted from forest to farmland during the last 1000 years. In the SMIC (1971) study, estimates are reported that over the last several thousand years, 1.5×10^7 to 2×10^7 km^2 of temperate forests have been converted to arable land and grassland, which corresponds approximately to the total extent of these terrain types in the temperate regions today. Sagan *et al.* (1979) adopts an estimate of 8×10^6 km^2, assuming that the present 1.2×10^7 km^2 of temperate evergreen and deciduous forest is 60% of the original forest area. Thus, the opinion of Sagan *et al.* that the assessments may be uncertain by a factor as large as 2 or 3 does not appear to be unduly pessimistic.

Our study considers neither the anthropogenic impact on arid regions, which may have caused a significant surface albedo increase (Otterman, 1977), nor the tropical deforestation which was analyzed by Potter *et al.* (1975). The study by Sagan *et al.* (1979) includes these two impacts. The 1 K decrease in temperature that they attribute to the anthropogenic effects over several millenia stems predominantly from the albedo changes in the arid, desert-fringe regions and in the tropical forest. Less than 0.1 K is attributable to deforestation in the nontropical regions, that is, to the difference in the winter albedo between forests and fields.

For purposes of discussion, we assume that the prehistoric nontropical forest extended over effectively two-thirds of the temperate and boreal regions in the Northern Hemisphere. The prehistoric climate is then at a point in the LA–HA interval one-third away from LA. By linear interpolation, the prehistoric hemispheric surface air temperature was 287.94 K. Assuming furthermore, as Sagan *et al.* (1979) did, that deforestation reduced the primordial forests by 40%, today's climate is placed in the LA to HA interval $0.6(2/3) = 0.4$ from HA. This corresponds to an interpolated hemispheric surface air temperature of 287.43 K and a reduction from the prehistoric times by 0.51 K. This reduction is much larger than the estimate by Sagan *et al.* (1979). The disparity in the two assessments stems in part from the fact that we consider the Northern Hemisphere only, with its extensive continents at the higher latitudes. Again by interpolation, the sensitivity of the hemispheric temperature to short perturbations in the solar constant increases since prehistoric times by a factor of 1.06 for a warming and by 1.13 for a cooling. The land temperature at 65°N decreases by 1.39 K and its sensitivity to perturbations of the solar constant increases by a factor of 1.11 for a warming and by 1.16 for a cooling. Our assessment of the prehistoric (PH) and the current (CR) climates within the interval LA–HA is presented in Table 3.

7. Conclusions

It emerges from this study that, by reducing the regional winter albedo, the temperate and boreal forests produce a warming effect on climate, especially at high latitudes. Since the forests reduce the albedo contrast between terrain with and without snow, the strength of snow–albedo feedback and hence, the sensitivity to perturbations, are weakened except in the case of a prolonged cooling perturbation at high northern latitudes. The location of the taiga/tundra ecocline is a significant climate-affecting and climate-reflecting parameter. The extent of deforestation and the location of the ecocline should be monitored, and their remote sensing from satellites is recommended.

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TABLE 3. A comparison of the current (CR) and prehistoric (PH) climates and the sensitivities to the solar constant in the LA–HA interval.

	LA	PH	CR	HA
	Fractional forest coverage south of ecocline			
	1.00	0.67	0.40	0.0
	Hemispherically averaged surface air temperatures and responses			
$1.00S_0$	288.58	287.94	287.43	286.67
$+2\%\ S_0$	+2.77	+3.00	+3.19	+3.47
	(1.00)	(1.08)	(1.15)	(1.25)
$-2\%\ S_0$	−2.67	−3.22	−3.66	−4.32
	(1.00)	(1.21)	(1.37)	(1.62)
	Land temperatures and responses at 65°N			
$1.00S_0$	269.35	267.61	266.22	264.13
$+2\%\ S_0$	+3.88	+4.53	+5.05	+5.83
	(1.00)	(1.17)	(1.30)	(1.50)
$-2\%\ S_0$	−3.96	−4.98	−5.79	−7.01
	(1.00)	(1.26)	(1.46)	(1.77)

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