Interaction between the Geobotanic State and Climate: A Suggested Approach and a Test with a Zonal Model

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ABSTRACT

A method is suggested for introducing long-term interaction between the geobotanic state and climate (a biogeochemical feedback mechanism) into climate models. It is based upon making the geobotanic state, characterized by the snow-free surface albedo and the water availability parameter, dependent upon the ratio of annual radiation balance to annual precipitation (the so-called radiative index of dryness). This approach is illustrated using a zonally averaged annual steady-state model which is based on the hemispheric climate model of Ohring and Adler. Zonal data statistics are employed to obtain simple relationships consistent with the zonality of the system. The heating parameterization of the original model is modified so that precipitation and cloud amount are computed using vertical velocity at 500 mb, which is calculated from the thermodynamic equation. Experiments with the model indicate that the simulated climate and geobotanic zones are in good agreement with observations. Sensitivity studies suggest that biogeochemical feedback has a negligible effect on the model’s response to solar constant variations but may be important in the evaluation of the long-term impact of surface albedo changes.

1. Introduction

Recently, a great deal of effort in climate modeling has been devoted to parameterizing the complicated feedback mechanisms which link the surface state to atmospheric processes. In current models, surface albedo and the water availability parameter (or evapotranspiration efficiency) appear to be the surface state parameters most frequently used in describing radiational and hydrological properties, respectively (see reviews by Carson, 1981; Eagleson, 1981). Commonly, these parameters are externally prescribed in those simple climate models which do not consider the subsurface hydrologic balance (e.g., Saltzman and Vernekar, 1971; Ohring and Adler, 1978).

To test climatic sensitivity to changes in surface state parameters, a number of numerical experiments with two- and three-dimensional climate models have been performed (e.g., Potter et al., 1979; see review by Mintz, 1982). However, the surface alterations have not been treated “within the context of a global climate feedback mechanism” (Cess, 1978), since the full effect of climate changes upon the surface state have not yet been considered. Indeed, since the surface state is closely related to climate, surface parameters should be included in a description of the long-term interaction between surface characteristics (soil, vegetation and moisture) and the atmosphere, depending on other dynamical variables of the earth-atmosphere system (Saltzman, 1967). The use of a model which includes such a biogeochemical feedback mechanism (which hereafter will be called “biofeedback”) could be useful for testing many kinds of climate modification, including those due to natural causes (e.g., Kutzbach, 1980) and those which are anthropogenically-induced (both intentionally and inadvertently) (e.g., Otterman, 1977; Sagan et al., 1979).

Numerical models of the biosphere, which embody the principal morphological and physiological factors, are being incorporated in some sophisticated climate models (Washington et al., 1979; Mintz et al., 1983). However, in simple climate models, the introduction of biofeedback employing explicit physio-biological processes remains infeasible (otherwise, such models would not be considered to be simple). An alternative approach is to make the surface state parameters (e.g., albedo and water availability) dependent on climatic variables. In terms of Lettau’s (1969) “climatonomic” principles, the surface state should be described by “response functions” (in our case, surface parameters) that are a physical consequence of a “forcing function” which could be a combination of climatic variables (precipitation, insolation, etc.).

Here, we attempt to effect this idea by employing a classification of climatic regions, observational data of surface albedo, and empirical relationships between

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2 In some experiments with general circulation models, water availability has been made a function of time-dependent (interactive) soil moisture. However, such treatments of the water availability use a prescribed soil water holding capacity. The latter can be identified with the vertical extent of the root zone which depends on the vegetation type. Therefore, the effect of a long-term climate change on possible vegetational modification is not taken into consideration.
components of the heat and water balances. Our major goal is to simulate the climate and geobotanic zones of the Northern Hemisphere. Unlike the "climatic atlas" constructed by Manabe (1975), which uses prescribed surface albedos and prescribed soil water holding capacity, we strive to simulate the geobotanic zones as a result of long-term surface-atmosphere interaction. In seems reasonable, as a first step, to test our parameterization with a zonally averaged climate model, which is simpler than three-dimensional models, and should be adequate for the purpose of this study (the simulation of zones).

The model we will use is a modification of the zonally averaged annual-mean hemispheric climate model of Ohring and Adler (1978), which will hereafter be referred to as OA. In addition to the simulation of the present climate and geobotanic zones, some experiments are also performed on the sensitivity of the model climate to solar constant variations and albedo modification.

2. Parameterization of biofeedback

a. General case

As a basis for the parameterization of biofeedback, we use the empirically determined relationship between geobotanic types and climatic conditions comprehensively described by Budyko (1974). According to Budyko, the principal geobotanic types can be subdivided using the so-called radiative index of dryness \( D^* \) defined as

\[
D^* = R^* / LP^*,
\]

where \( R^* \) is the mean annual surface radiation balance (net radiation), \( P^* \) the mean annual precipitation rate and \( L \) the latent heat of vaporization. (The asterisk is added in order to distinguish between the quantities used in the general (three-dimensional) case and those to be introduced later on for use in a zonal model.) His division of geobotanic types is shown in Fig. 1. The solid lines bound the domain of actually existing values of \( R^* \) and \( D^* \) (except for mountain regions) and the dashed lines distinguish the principal geobotanic types denoted by capital letters. In the case \( D^* < 0 \) (which can occur when \( R^* < 0 \)), the surface is assumed to represent arctic desert with permanent snow (Budyko, 1974).

As mentioned by Budyko (1974), in different latitudes lying within the limits of the same geobotanic zone, differences in the values of \( R^* \) correspond to certain geobotanic changes while the general character of vegetation remains unchanged. From Fig. 1, it can be seen that these changes are expressed in various types of woods and (to a lesser extent) grasslands, depending on \( R^* \).

In order to incorporate biofeedback into a climate model, we should try to relate the surface parameters \( \alpha^* \) (snow-free land surface albedo) and \( w^* \) (water availability parameter) to the geobotanic state characterized by \( D^* \).

Let us first derive a functional dependence \( w^* (D^*) \). Recall that when considering long-term annual climatology, \( w^* \) is defined as

\[
w^* = E^* / E^*_p\]

(2)

where \( E^* \) is the annual evapotranspiration rate and \( E^*_p \) the annual potential evapotranspiration rate (i.e., evapotranspiration from a fully wetted surface).

Budyko (1974) suggested that annual potential evapotranspiration for a land surface can be approximated by \( R^* \):

\[
E^*_p = R^* / L \quad (R^* > 0).
\]

(3)

This, in fact, is a basic assumption of Budyko's theory for using \( D^* \) as a climatic index. In his monograph,
he compared this approximation with other methods, emphasizing its advantage and convenience (see Budyko, 1974).

The annual surface water balance under steady-state conditions can be written as

\[ M^* = P^* - E^* , \]

where \( M^* \) is the annual runoff. Division by \( P^* \) and rearrangement yields

\[ E^* = (1 - C^*) P^* , \quad (4) \]

where \( C^* = M^*/P^* \) is the runoff ratio.

Analysis of empirical data on components of the heat and water balances indicates a functional interdependence of \( C^* \) and \( D^* \) (Budyko, 1974). It should be noted that these concepts introduced by Budyko have been used in some diagnostic studies of regional climate (Lettan, 1969; Hare, 1977; Kutzbach, 1980).

Employing the empirical relationship suggested by Lettun (1969),

\[ C^* = 1 - \frac{\tan h D^*}{D^*} , \quad D^* \geq 0 , \quad (5) \]

together with Eqs. (1)-(4), we obtain

\[ w^* = \frac{\tan h D^*}{D^*} , \quad D^* \geq 0 . \quad (6) \]

When \( D^* \) is negative, \( w^* \) is not used in parameterizing the latent heat flux and, therefore, need not be specified. Note that in order to be consistent with (3), a model which employs (6) should calculate evapotranspiration as

\[ E^* = w^* R^*/L , \quad R^* \geq 0 . \quad (7) \]

It should also be noted that other expressions for \( w^* (D^*) \) can be obtained by combining (3) with any of the known empirical relationships between the components of the hydrologic balance (see review by Eagleson, 1981). The behavior of \( w^* (D^*) \) would be similar for all such relationships.

We stress that since the water availability parameter can be expressed as a single-valued function of \( D^* \), \( w^* \) itself can be used as a climatic index in agreement with Sellers (1965). Note that the differences between the values of radiation balance for moistened and actual surfaces (see footnote 3) may be substantial in arid regions (i.e., when \( D^* \) is large). From the structure of \( w^* (D^*) \), it can be seen that with increasing \( D^* \) the changes in \( w^* \) become very small. Therefore, the error in \( w^* \) due to ambiguity in specifying \( D^* \) would be negligible. It also follows from Eq. (7) that in this case the absolute error in estimating \( E^* \) would be much less than the possible error in \( R^* \). (On the other hand, small changes in \( w^* \) lead to large differences in \( D^* \), indicating that \( D^* \) is preferable as a general purpose climatic index.)

Turning to a consideration of the snow-free surface albedo \( \alpha^* \), we note that unlike \( w^* \), values for this parameter are available from observations. This facilitates the construction of a functional dependence of \( \alpha^* \) upon geobotanic type. Such a relationship could be an approximation of compiled data on the albedo of natural surfaces (e.g., see Hummel and Reck, 1979; Budyko, 1974). Observations indicate a general tendency for \( \alpha^* \) to increase with transition from forests to arid lands. Within the Forest zone (\( D^* < 1 \)), the dependence of albedo upon forest type (characterized by different \( R^* \)) (see Fig. 1) should be considered. Hence, a more exact formulation for \( \alpha^* \) would make it a function of both \( D^* \) and \( R^* \).

Thus, incorporation of long-term geobotanic state-climate interaction into three-dimensional climate models potentially can be made by employing relationships for \( \alpha^* (D^*, R^*) \) and \( w^* (D^*) \), together with other equations of the system provided that the annual radiation balance and precipitation are available from the model calculations at each grid point.

b. Zonal case

Our goal is to obtain a consistent set of equations containing climatological quantities zonally averaged over latitude belts. For further derivations we will employ zonal data from Budyko (1978, Tables 5.1 and 5.3, hereafter referred to as BT), using the values given for 10° latitude belts at the middle of each belt. Subscripts \( l \) and \( oc \) will imply averaging over land and ocean, respectively, at each latitude belt. The quantities \( R, E \) and \( P \) denote radiation balance, evapotranspiration and precipitation, respectively, zonally averaged over the earth as a whole in each latitude belt. Here, \( P \) and \( E \) can be represented as

\[ P = IP_l + (1 - l) IP_{oc} , \]

\[ E = IE_l + (1 - l) IE_{oc} , \quad (8) \]

where \( l \) is the fraction of land in each latitude belt.

Direct application of the general approach is obstructed because of the nonlinearity of (1) and (6) and the nonhomogeneity of \( D^* \) along some latitudes (especially in the subtropics). Therefore, for use in a zonal model, we introduce by analogy with \( D^* \) a "zonal" index of land dryness defined as

\[ D = R_l / LP_l . \quad (9) \]

It is obvious that this ratio could differ from a value of zonally averaged \( D^* \) at a latitude with diverse land surface types. Nevertheless, in view of the predominantly zonal character of vegetational distributions (Walter, 1973), (9) might be employed for specifying the predominant latitudinal land surface type within the confines of a zonally averaged model.

Considering the above, the introduction of \( D \) requires a division of geobotanic types different from that of Budyko (Fig. 1). Such a division can be made by using a "vegetational map" (see, e.g., Walter, 1973) as follows. First, from such a map we visually locate the boundaries between geobotanic zones. Obviously, these locations are determined rather subjec-
tively. (For example, a certain difficulty arises when specifying boundaries between desert and semidesert). However, this is of little importance since we will be interested mainly in relative changes of climate and zones (e.g., as a result of changes in boundary conditions). Then, we interpolate (by cubic spline) between the values of $D$ calculated from (9) using BT. Finally, the values of $D$ corresponding to the boundaries between the principal geobotanic zones are found from the latitudinal $D$-distribution. The suggested division according to $D$ is shown in Table 1, together with the original division by $D^*$. Grassland includes steppe, prairie, savanna, meadows, etc. The Arctic surface type exists whenever $R_i < 0$ (or, equivalently, $D < 0$). The land surface in this case may still be only partly covered by snow.

In Table 1, the $D$ limits for $D > 1$ are lower than those of $D^*$. This is the consequence of using $R_i/LF_i$ (which is a ratio of zonally averaged land values) instead of averaging $D^*$ along latitudes in which areas with very little precipitation alternate with areas of heavy rainfall. Such a "smoothing" effect is consistent with the framework of simple zonal models.

Furthermore, by analogy with $w^*$ we denote

$$w = \frac{E_i}{(E_p)_i},$$

(10)

which can be interpreted as "zonal land" water availability and may differ from a value which would be obtained by averaging $w^*$ over land. Zonal averaging over land for Eq. (3) yields

$$(E_p)_i = \frac{R_i}{L_i}, \quad R_i > 0,$$

(11)

and using (10) we obtain

$$w = \frac{LE_i}{R_i},$$

(12)

A formula for zonal evapotranspiration $E$ containing the introduced parameter $w$ can be obtained using (8):

$$E = \frac{R}{L} [lwa_2 + (1 - l)w_0a_1], \quad R > 0,$$

(13)

where

$$w_0 = \frac{LE_{oc}}{R_{oc}}, \quad a_1 = \frac{R_{oc}}{R}, \quad a_2 = \frac{R_j}{R}.$$

(14)

In the case $R \leq 0$, we assume $E$ equal to 0. This can be justified by the relatively small values of evaporation in the polar regions (Budyko, 1974).

![Fig. 2. Correlation between observed mean latitudinal values of radiation balance over ocean $R_{oc}$ and over Earth as a whole $R$ at different latitude belts for the Northern (dots) and Southern (circles) Hemispheres.](image)

The quantities introduced, $w$ and $w_0$, represent the fractions of the radiation balance which are used for latent heat flux over land and ocean, respectively. Their specification is described below. Note that actual evaporation at the ocean surface is equal to its potential value with $w^*$ equal to 1. However, since (3) may not be valid for ocean, the ratio $w_0$ can be different from 1. (This is due to the additional term in the heat balance equation which accounts for ocean currents).

Let us show that $a_1$, $a_2$ and $w_0$ can be taken constant and determined from the zonal data statistics (BT). Consider, for example, the relationship between $R_{oc}$ and $R$. As can be seen from Fig. 2, the ratio of $R_{oc}$ to $R$ is approximately the same at all latitudes, i.e., for different climatic conditions. Therefore, we can assume that for any latitude the relationship between $R_{oc}$ and $R$ also would hold during possible climatic changes.

Analogous arguments can be suggested with respect to $a_2$ and $w_0$, in addition to

$$a_3 = \frac{P_{oc}}{P_i}$$

(15)

which is used below. All the above ratios can be assumed constant with a sufficient degree of accuracy. They are determined by least-squares fit and given in Table 2. To illustrate, in Fig. 3 we compare the observed values of $R_{oc}$, $R_i$, $LE_{oc}$ and $LP_{oc}$ (from BT) with those computed by the formulas $R_{oc} = a_1R$, $R_i = a_2R$, $LE_{oc} = w_0R_{oc}$ and $LP_{oc} = a_3LP_i$, respectively.

Table 1. Division of land surface types by $D^*$

<table>
<thead>
<tr>
<th>Surface type</th>
<th>Arctic</th>
<th>Tundra</th>
<th>Forest</th>
<th>Grassland</th>
<th>Semi-desert</th>
<th>Desert</th>
</tr>
</thead>
<tbody>
<tr>
<td>$D^*$</td>
<td>0</td>
<td>0.33-1</td>
<td>0.33-1</td>
<td>1-2</td>
<td>2-3</td>
<td>&gt;3</td>
</tr>
<tr>
<td>$D$</td>
<td>0</td>
<td>0.33-1</td>
<td>0.33-1</td>
<td>1-1.4</td>
<td>1.4-1.7</td>
<td>&gt;1.7</td>
</tr>
</tbody>
</table>
for each latitude. Figure 3 indicates that these approximations describe the observational data with rather high correlation coefficients.

Note that in Fig. 2 the values of the $R_\infty$ to $R$ ratio for the Southern Hemisphere fit the regression obtained for the values in the Northern Hemisphere surprisingly well. Moreover, the ratios $a_2$ and $w_0$ can also be taken constant for the Southern Hemisphere (with slightly different values), justifying the assumption concerning the relationships in (14). However, this is not true for $a_4$, which varies greatly in the Southern Hemisphere. Therefore, one should seek a relationship different from that in (15) in order to apply our approach to simulate "zonal" climate in the Southern Hemisphere.

In Fig. 4, dots represent plotted values of $w$ [obtained from (12) using BT] versus values of $D$ [obtained from (9) using BT]. A good correlation can be seen between $w$ and $D$. Since we do not have enough data concerning small and large $D$, we try to employ the relationship $w^*(D^*)$ derived earlier and shown in the same figure. In other words, we assume that the character of the functional dependence $w(D)$ remains analogous to that in (6). Regression between $w$ and $w^*$ yields a good correlation with a coefficient $r = 0.987$.

Finally, the suggested approximation is

$$w = 1.23 \frac{\tanh D}{D} - 0.33, \quad D \geq 0. \quad (16)$$

Since $w$ cannot become negative, the upper limit for $D$ is 3.7. This value is unlikely to be reached in a zonal model. However, whenever $D$ exceeds this value, $w$ should be set to 0. This relationship is shown by the dashed curve in Fig. 4.

Another parameter which we intend to make dependent upon geobotanic state is zonally averaged snow-free surface albedo $\alpha$. In Fig. 5, values of $\alpha$ from Sellers (1973) are plotted versus $D$ (Fig. 5a) and $R_l$ (Fig. 5b) taken from BT. It can be seen that a correlation between $\alpha$ and $D$ exists only for $D > 1$ (dots), while for $D < 1$ we note a correlation between $\alpha$ and $R_l$ (circles). This corresponds to the marked dependence of forest types (having different albedos) within the Forest zone upon radiation balance, as mentioned earlier.

Employing (14), $\alpha(R_l)$ can be expressed through $R$. In the area of negative radiation balance corresponding to $D < 0$ [see Eq. (9)], data on values of $R_l$ are not available. Therefore, when $D < 0$ we prescribe $\alpha$ as a function of latitude using Sellers' (1973) data.

Thus, the suggested approximation can be presented as

$$\alpha = \begin{cases} 0.07 + 0.06D, & D \geq 1 \\ 0.20 - 0.002a_2R, & 0 < D < 1 \\ \alpha(\phi), & D \leq 0, \end{cases} \quad (17)$$
where $\phi$ is latitude and $R$ is expressed in W m$^{-2}$. Note that the discontinuity in $\alpha$ at $D = 1$ in (17) does not introduce additional difficulties in the computations.

Even with so few available data points, the suggested parameterization grasps the general trend of changes in $\alpha$. Indeed, since for $D > 1$ (i.e., for areas with insufficient moisture) an increase in $D$ corresponds to transition toward "aridity," an increase in $\alpha$ with $D$ implies that dryer surfaces have higher albedos. On the other hand, for $D < 1$ (i.e., for the Forest zone) the tendency of $\alpha$ to increase with latitude (lower radiational balance) is seen in empirical data on forest albedos at different latitudes (Posey and Clapp, 1964).

Thus, we have expressed $\alpha$ and $w$ through $D$, which in turn can be calculated using zonal climatic variables $R$ and $P$. Indeed, combining (8), (9), (14) and (15), we obtain

$$D = \frac{R}{LP} a_2 [l + (1 - l) a_1].$$  \hspace{1cm} (18)

Introduction of (16) and (17) together with (18) into a zonal model implies that land-surface state is not prescribed but varies with climate while the geobotanic type (characterized by $D$) becomes a new predicted variable.

Summarizing this section, we note that the use of the proposed parameterization is restricted to simulations of climatic changes in the Northern Hemisphere with the present land-ocean distribution. Geobotanic changes occur over considerably shorter periods than the relaxation times of processes in the cryosphere or changes in shape and location of the land masses. Hence, in spite of the simplified assumptions made here, the proposed parameterization may be usefully applied in a zonal model for simulating climatic impacts of surface alterations in the Northern Hemisphere which have occurred during the past thousand years or will occur within a similar future period as a result of modifications in the biosphere.

3. The model

We will present below only a short summary of the structure of the OA model, except for the changes that we have introduced which are given in detail. The full description of the basic model can be found in OA.

a. Basic model

The dynamical model is described by a system of zonally averaged quasi-geostrophic potential vorticity equations for two levels in the atmosphere, 250 and 750 mb. These equations are obtained by combining the vorticity and thermodynamic energy equations (e.g., see Sela and Wiin-Nielsen, 1971). The other governing equations of the model are the diagnostic equation for the thermal streamfunction and the surface heat balance equation.

Atmospheric and surface heating processes are parameterized as in OA except for latent heat flux at the surface and latent heat release in the atmosphere, which are described later in this section.

b. Computation of vertical velocity $\omega$ at 500 mb

Although $\omega$ figures implicitly in the dynamical system, it was not computed in OA. However, our modification of the OA heating model, which appears later in this section, employs an empirical relationship
between the water balance and $\omega$. Therefore, following Sela and Wiin-Nielsen (1971), we compute $\omega$ from the thermodynamic equation at the 500 mb level,

$$
\omega = \frac{2f_0}{\sigma P_2} \left[ \frac{\partial \psi_T}{\partial t} - \frac{1}{a^2 \cos \phi} \frac{\partial}{\partial \phi} \left( K_H \cos \phi \frac{\partial \psi_T}{\partial \phi} \right) \right] - \frac{R_0}{2c_p f_0} H_a,
$$

where $\psi_T$ is the zonally averaged thermal streamfunction, $f_0$ the Coriolis parameter at 43°N, $c_p$ the specific heat at constant pressure, $\sigma$ is a standard value of the static stability at 500 mb, $R_0$ is the gas constant, $P_2 = 500$ mb, $t$ is time, $a$ is the radius of the earth, $\phi$ is latitude, $H_a$ is the zonally averaged rate of heating of the atmosphere, and $K_H$ is the exchange coefficient for the meridional transport of sensible heat at the 500 mb level.

In deriving (19), it is assumed that the meridional transport of sensible heat can be parameterized by analogy with diffusive-type treatment of the meridional transport of quasi-geostrophic potential vorticity (see Sela and Wiin-Nielsen, 1971).

The values of $\omega$ are computed for each grid point with the heating processes lagging by half a time step, consistent with the finite difference form of other prognostic equations of the basic model:

$$
\omega^{i+1/2,j} = \frac{q^2 P_2}{f_0} \left( \frac{\psi_T^{i+1,j} - \psi_T^{i,j}}{\Delta t} - \frac{ADV^{i+1,j} + ADV^{i,j}}{2} - \frac{R_0}{2c_p f_0} H_a^{i,j} \right),
$$

where $i$ is the time index, $j$ the grid point index and

$$
q^2 = \frac{2f_0}{\sigma P_2},
$$

which is taken constant and modified in the course of the tuning experiments from the value of 4 (used by Sela and Wiin-Nielsen, 1971, as well as OA) to 2.5 (used by Wiin-Nielsen, 1972). The advection term is computed as

$$
ADV^{i,j} = \frac{1}{a^2 \cos \phi)^2} \left[ (K_H \cos \phi)^{j+1/2}(\psi_T^{i,j+1} - \psi_T^{i,j}) - (K_H \cos \phi)^{j-1/2}(\psi_T^{i,j} - \psi_T^{i,j-1}) \right],
$$

where $\Delta \phi$ is the horizontal size.

We note here that, working with a mean annual model, we seek a stationary solution, i.e., a climate that is in equilibrium with the external forcing. Therefore, integration of the equations containing time derivatives is just a method of computation (where $\Delta t$ is an iterative parameter) we use to obtain this equilibrium climate.

The values of the latitudinally dependent coefficients $K_H$ for the band 80–30° were taken from Wiin-Nielsen and Sela (1971), except for the value at 30° which was adjusted so that $\omega$ at this latitude would be closer to the zonal mean $\omega$ given by Oort and Rasmussen (1971). The same principle was used for assigning the $K_H$ values at 20 and 10°, which were absent in Wiin-Nielsen and Sela (1971). The latitudinal distribution of $K_H$ is shown in Table 3. At the boundaries (0 and 90°), $K_H$ are not employed in the finite difference scheme, hence values are given only in the band 10–80°N. The negative value at 10° can allow for the boundary effect in a hemispheric model. It can also be related to the fact that a two-level model does not take into account the counter-gradient heat transport at high elevations (see Wiin-Nielsen and Sela, 1971). Actually, the above adjustment of $K_H$ in low latitudes implies tuning for better simulation of the Hadley cell. We stress that this adjustment and further modifications of the model have been made not to improve the original model but to make it applicable for our purposes.

c. Changes in the OA heating model

Our modifications to the parameterizations of OA include the new parameterization of evapotranspiration [Eq. (13)] which is adopted for consistency with the suggested biofeedback parameterization (Section 2), and a new precipitation/cloudiness approximation using $\omega$ which is discussed below.

The heating of the atmosphere due to condensation processes is parameterized in OA by

$$
LP = LE + g(N - \overline{N}),
$$

where $N$ is the fractional cloud amount at a particular latitude belt, $\overline{N}$ is the average annual fractional cloud amount in the Northern Hemisphere and $g$ is a constant independent of latitude. Actually, it is assumed in OA that the cloudiness deviation $(N - \overline{N})$ is proportional to the flux convergence of water vapor through the atmospheric circulation. Since the cloudiness is prescribed in OA, this flux convergence is also prescribed and can be regarded as one of the external parameters (or boundary conditions) of the OA model.

Since changes in precipitation can result from changes both in surface hydrology and atmospheric dynamics, the use of (20) is not suitable for our purposes (Section 1) unless the cloudiness is allowed to vary with the dynamics.

| Table 3. Latitudinal distribution of exchange coefficients $K_H$ (10° m² s⁻¹) for sensible heat at 500 mb. |
|--------|--------|--------|--------|--------|--------|--------|--------|--------|
| $\phi$ (deg) | 10 | 20 | 30 | 40 | 50 | 60 | 70 | 80 |
| $K_H$ (10° m² s⁻¹) | -0.6 | 0.6 | 1.5 | 1.8 | 3.3 | 3.8 | 2.3 | 2.0 |
Recall that in many of the experiments on surface alterations (e.g., desertification experiments) there have been attempts to relate changes at the land surface to changes in atmospheric dynamics which then result in modification of precipitation patterns. Since our ultimate goal is to apply our model to this kind of experiment, we have sought a relationship between surface hydrology and the dynamical behavior of the atmosphere, with the relevant features of the atmospheric dynamics represented by $\omega$. For this, we assume the linear relationship

$$P - E = -a_4 \omega \quad \text{for} \quad \phi < 70^\circ,$$  \hspace{1cm} (21)

where $a_4$ is a constant obtained by least-squares fit using the annual mean zonal $\omega$ from Oort and Rasmusson (1971) and values of $(P - E)$ from BT. This relationship is obviously not valid for the polar region where $\omega > 0$ and $E$ is negligible. In such a case, according to (21), $P$ would become negative. Due to this and the absence of data on $P$ and $E$ for latitudes north of $70^\circ$, for this area we use the old parameterization expressed by (20). Thus, the parameterization of the latent heat release in the atmosphere is taken as

$$LP = \begin{cases} LE - L_a \omega, & \phi < 70^\circ \\ LE + g(N - \bar{N}), & \phi > 70^\circ. \end{cases}$$

(22)

Since hemispherically averaged $\omega$ is zero in our system, if (21) was in force for the whole hemisphere, the total hemispheric rainfall would equal the total hemispheric evapotranspiration. This is not maintained in (22) but the error is less than 4% of the hemispheric average values of $P$ and $E$.

Essentially, this parameterization for $\phi < 70^\circ$ implies that zonal mean subsidence at 500 mb ($\omega > 0$) leads to divergence of the water vapor flux while zonal mean upward motion ($\omega < 0$) results in its convergence.

Although cloudiness $N$ does not explicitly appear in our biofeedback parameterization, its influence is felt indirectly through $R$ since changes in cloudiness lead to changes in the radiational regime. We believe that since change in precipitation is a physical consequence of change in cloud amount, it is natural to introduce an approximation for cloudiness through $\omega$ consistent with that of precipitation. A parameterization for $N$ (for $\phi < 70^\circ$) can be obtained as a combination of (20) and (21). For the area north of $70^\circ$N where $\omega$ is positive, such a relationship between $N$ and $\omega$ is not valid (observational data show an increase in $N$ in the polar region). Therefore, we prescribe cloudiness for this region as in OA. Thus we parameterize $N$ as

$$N = \begin{cases} \bar{N} - a_5 \omega, & \phi < 70^\circ \\ \frac{N(\phi),}{N(\phi),} & \phi > 70^\circ, \end{cases}$$

(23)

where $a_5 = a_4 L / g$. The small imbalance that arises from the separate formulations for $\phi < 70^\circ$ and $\phi > 70^\circ$ can be neglected for the reason given above.

The physical sense of (23) is obvious: downward vertical motion reduces the cloud amount, while upward motion increases the cloud amount relative to the average. We note, however, that the hemispheric average cloud amount $\bar{N}$ is still an external parameter which must be prescribed. In this model, we use $\bar{N} = 0.51$, which is taken from London (1957) and was used in OA.

A comparison of the proposed approximations for $P - E$ and $N - \bar{N}$ with observations is shown in Fig. 6.

Although cloudiness is allowed to vary, the limitations of the model with respect to the description of humidity should be borne in mind. (Surface relative humidity and the shape of the vertical profile of specific humidity are prescribed; see OA.)

d. Surface characterization

The earth's annual mean surface reflectivity is computed as (in OA notation)

$$r(\phi) = \sum_{j=1}^{4} \sum_{i=1}^{3} f_i(\phi) n_{ij}(\phi) r_{ij}(\phi),$$

where $i$ is the index for three possible surface types (open ocean, land snow/ocean ice and snow-free land), $j$ the index for season, $r_{ij}(\phi)$ the reflectivity of surface type $i$ at latitude $\phi$ during season $j$, $n_{ij}(\phi)$ the fraction of latitude belt covered by surface type $i$ during season $j$, and $f_i(\phi)$ the weighting function accounting for the seasonal change in solar radiation. The values of $f_i(\phi)$, $n_{ij}(\phi)$ and $r_{ij}(\phi)$ are prescribed as in OA. In our notations, $r_{ij}$ corresponds to $\alpha$, and when biofeedback is included it is computed as a function of $D$ and $R$ according to (17). The actual

![Fig. 6. Correlation between observed and computed $P - E$ (dots) and $N - \bar{N}$ (circles).](image-url)
incorporation of biofeedback is accomplished by computing $D$ from (18) and using (17) and (16) for specifying $\alpha$ and $w$, respectively.

Computation of $D$ permits the specification of the geobotanic type at each grid point according to the division given in Table 1. While in the no biofeedback model such a procedure implies a diagnosis of vegetational zones, the model including biofeedback contains the geobotanic type as a predicted variable interacting with the climate.

In concluding this section, let us summarize the modifications we have introduced in OA:

1) Computation of vertical velocity ($\omega$) at 500 mb has been added for use in the new parameterization of precipitation;
2) The parameterization of evapotranspiration has been modified to be consistent with the surface characterization;
3) The water vapor flux convergence (precipitation minus evapotranspiration) has been made a function of $\omega$ to allow for the dynamical factor in the parameterization of precipitation;
4) The cloudiness distribution has been allowed to vary with $\omega$ to be consistent with the above;
5) We have made $\alpha$ and $w$ predicted variables while geobotanic type (specified by $D$) is also predicted, thus incorporating the geobotanic state–climate interaction.

These modifications are shown schematically in Fig. 7.

4. Results

The first objective of this study was to attempt to simulate both the observed climate and the geobotanic zones. Our second objective was to carry out some experiments on the sensitivity of climate to solar constant variations and albedo modification, and then to compare the results with those obtained by other investigators.

In order to clarify the impact of biofeedback incorporation, we had to start with the no-biofeedback version of the model. Therefore, we initially introduced all of the previously-described modifications (Section 3) into the OA model, keeping $\alpha$ and $w$ prescribed (using data from Sellers, 1973, and BT) and then tuned the model. The tuning process was similar to that described in OA. During the course of the tuning experiments some of the input parameters were modified from those used in the original OA model. The cloud particle optical depth $\tau$, the value of $c$ in the parameterization of the sensible heat flux and the value of $h$ in the oceanic flux parameterization were adjusted for better simulation of the mean surface temperature, the vertical temperature gradient and the meridional temperature distribution, respectively. The value of $q^2$ (see Section 3) was adjusted so that winds would be closer to those observed. These parameters, together with other new constants used in the model, are listed in Table 2. In addition, the three values of the $K_{fH}$ coefficients in low latitudes were adjusted to obtain a better simulation of $\omega$ (see Section 3). (The process of tuning, and its purpose and justification, have been outlined in OA.)

The tuning was completed when we were able to reproduce the present equilibrium climate to the extent that temperatures and winds generally agreed with those presented in OA as well as with observational data. Certain discrepancies lay within the range of errors expected in such models: 1–2°C for temperature and 1–2 m s$^{-1}$ for wind speed, which are probably close to the uncertainty limits for the observed values themselves. For example, a realistic value of 15°C was obtained for the annual mean hemispheric surface temperature.

After we were satisfied with the simulated climate, we started experiments with biofeedback included. Without any additional tuning, we found that the results of these experiments had very small deviations when compared with those obtained with the no biofeedback version of the model. As would be expected, differences were found mainly in surface temperatures ($\sim0.5°C$) and in heat fluxes at low latitudes. If the meridional profiles of $D$ and $R$ had been exactly equal to the observed profiles and if our $\alpha$- and $w$-parameterizations were perfect, we would have obtained exactly the same equilibrium climate as in the case with $\alpha$ and $w$ prescribed (in OA this point is discussed with respect to ice–albedo feedback).

Since the difference in results with and without biofeedback was insignificant, here we have chosen to present only the results obtained with the model including biofeedback. These are compared with the available mean annual zonal data. This experiment will hereafter be referred to as the control experiment.

a. Simulation of observed climate and geobotanic zones

Computed values of winds and temperatures are shown in Figs. 8 and 9. The observations are taken from Oort and Rasmussen (1971). Simulation of these quantities was one of the goals in OA and was

![Fig. 7. Schematic diagram showing modifications introduced to the OA model.](image-url)
discussed there in detail. The latitudinal profiles of both winds and temperatures are simulated with the same degree of success as in OA except for some minor differences (such as the underestimation of the peak westerlies at the 750 mb level and of 500 mb temperatures north of 40°N, with the maximum deviations of the computed temperature from the observed less than 4°C). Therefore, rather than dwell on the analysis of the above results, we will devote our attention to a discussion of the quantities which were not presented in OA but which play an important role in our model.

Figure 10 shows the simulated 500 mb vertical velocities and the cloud amount \( N \). The model calculations of \( \omega \) are in reasonable agreement with those of Oort and Rasmusson (1971) which are based on wind observations. This can be partly explained by the adjustment of the \( K_{H} \)-coefficients in low latitudes. The model succeeded in simulating the following realistic features of the atmospheric circulation: the ascending and descending branches of the Hadley cell, the broad subsidence zone in the subtropics, and the convergence zone in temperate latitudes.

As far as cloudiness is concerned, the fairly close agreement with observations can be attributed to the simulation of \( \omega \), since the model cloud distribution depends solely upon the \( \omega \) distribution. The maximum deviation of the computed \( N \) from the observed is found near the equator and is consistent with the largest difference between our approximation of \( N - \overline{N} \) and London's (1957) data in the latitudinal belt 0–10°N (see Fig. 6).

The components of the computed and observed heat and hydrologic balances are given in Figs. 11 and 12. The observed data are taken from BT.
From the viewpoint of vegetation, the active factors of climate are evapotranspiration, potential evapotranspiration and precipitation (Mather and Yoshioka, 1967). Therefore, it is of crucial importance that these parameters be adequately simulated in a climate model in which the surface is characterized by a combination of these same parameters.

Unlike the OA model, in which simulation of the sensible and latent heat fluxes at the surface and in the atmosphere was not a primary objective and therefore not presented, we considered it particularly important to reproduce the heat components as accurately as possible because of their influence on our surface characterization.

Figure 11 shows that the agreement of the model fluxes with observations is fairly close. The heat income due to ocean currents represents the smallest of the four heat components, so that, for the most part, the sum of the sensible and latent heat fluxes compensates the net radiative heating. The maximum sensible heat flux is found in the subtropics where it has its largest contribution in the land surface heat balance.

In Fig. 12, the two precipitation maxima (at the equator and in the temperate latitudes) and the two minima (in the subtropics and the polar region) coincide with their observed locations. Consistent with our parameterization of the flux convergence of water vapor, the maximum negative value of simulated \( P - E \) is at the latitude of the strongest sinking motion (\( \omega > 0 \)) and the minimum cloud amount (see Fig. 10).

As shown by Fig. 13, this region is characterized by the maximum \( D \). In this figure, we present the computed latitudinal distribution of \( D \) along with that derived from BT. Although the shape of the model latitudinal distribution is similar to the observed, the model values are slightly lower. This implies that the model surface state is somewhat "wetter" than observed. In the subtropics, this can be partially accounted for by an overestimate of the precipitation. In temperate latitudes, it is a consequence of the underestimated radiation balance. Nevertheless, in our model these differences in \( D \) are sufficiently small to produce a rather successful simulation of the surface state parameters \( \alpha \) and \( \omega \). In Fig. 14, this simulation is compared with Sellers' (1973) values for snow-free land surface albedo and with values of water availability calculated from BT [using (12)]. For the area with \( D \) greater than 1 (between 15° and 45°N), the latitudinal variation of water availability is in antiphase with that of land surface albedo. The high values of \( \omega \) near the equator and in temperate latitudes reflect the fact that the evapotranspiration in the forest zone approximates its potential value.

Figure 15 shows the computed (right) and BT (left) distributions of the geobotanic zones. The idealized continent–ocean system is similar to that of Sellers (1973) except for the symmetry of the land masses with respect to the equator–pole axis. In each 10°
latitude belt, land occupies its presently-observed fraction of the belt. The solid latitudinal lines separate the geobotanic zones; their locations are established using the computed and observed latitudinal distributions of $D$ from Fig. 13 and the $D$ division of Table 1.

The transitions from Forest to Grassland in the tropics and temperate latitudes are at about the same latitudes as observed. The model produces a Desert zone in the region of maximum subsidence between 20° and 30°, which is in agreement with the “desert core” in the real world. The southern margin of the Desert zone (between 18 and 23°) corresponds to the Semidesert zone where desertification phenomena seem most widespread.

Figure 15 shows that the general features of the distribution of predominant zones in each latitude belt are simulated by the model and agree with those obtained from geobotanic studies (e.g., Walter, 1973).

Thus, our results indicate that, on the whole, the simulated climate and geobotanic zones are in good agreement with observations.

b. Sensitivity experiments

In addition to the simulation of the observed climate and geobotanic zones, the model has been further checked by conducting some sensitivity experiments and comparing the results with those obtained by other investigators in similar experiments. These include tests on climate sensitivity to solar constant variations (manifested by changes in the mean surface temperature) as well as a so-called desertification test in which the climatic response to an increase in surface albedo has been investigated.

Here our major goal is to show that the sensitivity of the computed climate is comparable with that in other studies. Besides testing the model response against results obtained with other models, we can tentatively examine the effect of incorporating biofeedback. For this, each experiment consisted of two runs, one including and the other not including biofeedback. In the latter case, the prescribed latitudinal distributions of $\alpha$ and $\omega$ were taken from the control experiment.

It has been shown by a number of climate studies (e.g., Lian and Cess, 1977) that climate sensitivity depends, among other things, on the ice-albedo feedback mechanism, which enhances the model response to changes in the boundary conditions. Since it is of interest to determine the impact of biofeedback alone, as well as in combination with ice-feedback, we carried out experiments both with and without ice feedback. Our ice-feedback parameterization was taken from OA (in which the seasonal land snow/ocean ice covers and snow/ice albedos are linearly related to the annual mean surface temperatures).

In this series of experiments, the results of each run were compared to the equilibrium state obtained in the control experiment discussed earlier (simulation of observed climate and geobotanic zones).

1) Sensitivity to solar constant variations

The hemispheric mean surface temperature change (perturbed minus control) $\Delta T_s$ has been evaluated by running our model with two “perturbed” values of the solar constant, $0.99 S_0$ and $1.01 S_0$, where $S_0 = 1360 \text{ W m}^{-2}$ is the current solar constant used in the control experiment.

A comparison of the present results with those obtained in OA is shown in Table 4. It can be seen that the modifications introduced in the original OA model do not significantly change the sensitivity to solar constant variations. The results for both runs, with and without biofeedback, were virtually the same. The differences between the temperature changes at all latitudes for these two runs were
sufficiently small (~0.1°C) to produce no noticeable difference in \( \Delta T \). Thus, we conclude that the incorporation of biofeedback, in the form in which we have parameterized it, hardly affects the sensitivity of the computed zonal climate to solar constant variations.

Another estimate of the sensitivity of climate to changes in the solar constant with biosphere–albedo feedback was made by Cess (1978). That study considered only one effect of vegetational modification, albedo change, which was assumed proportional to the change in the zonal surface temperature. With that kind of biosphere–albedo feedback included, Cess estimated an increase of the sensitivity by a factor of 1.8.

Surface temperature alone is probably not the best descriptor of a climatic region (and hence of vegetational cover). A combination of the heat and moisture balance components (which in our model is \( D \)) is preferable for this purpose (Mather and Yoshioka, 1967; Budyko, 1974).

In order to obtain results somewhat comparable to Cess’, we have also run our model with an albedo–temperature feedback in a form similar to that described in his paper. As a result, we obtained a ~40% increase in sensitivity. We consider therefore that the 80% increase reported by Cess can probably be attributed to his method of parameterizing biofeedback.

The results of our experiments indicate that although solar constant variations do modify surface temperatures, the relative changes in radiation balance and precipitation can be of about the same magnitude and have the same sign. The net result is a small change in \( D \) to which \( \alpha \) and \( w \) are not sensitive. This effect can be interpreted as weak sensitivity of vegetational cover to those climatic changes (in terms of temperature) which would certainly be “felt” by human beings.

### 2) Sensitivity to Changes in Land Surface Albedo

As another check on the model, we carried out an experiment similar to that performed by Ellsaesser et al. (1976) with the zonal atmospheric model of the Lawrence Livermore National Laboratory (LLNL).

A value of 0.35 for land surface albedo was assigned in the belt between 10 and 20°N and held constant during the experiment. This value is the same as that used by Ellsaesser et al. [and originally by Charney et al., 1975, in their desertification test with the Goddard Institute for Space Studies (GISS) GCM].

For this experiment, we would have preferred that the albedo not exceed the maximum “zonal land” surface albedo of 0.18 in Sellers’ (1973) data, in order to be consistent with the zonality of the system. However, the high value of 0.35 is used in order to have our test resemble that of the LLNL model. For the same reason, in perturbation region (between 10 and 20°) we kept \( w \) fixed, equal to the value for the control experiment, in order to eliminate the influence of changes in evapotranspiration due to this factor. Thus, we attempted to reproduce with maximum similitude the prescribed perturbation used in the LLNL model. In the run with biofeedback, \( \alpha \) and \( w \) were prescribed only in the perturbation area. When biofeedback was not included, these parameters were prescribed everywhere.

Changes (perturbed minus control) in precipitation and temperature for both runs (with and without biofeedback) were compared with those obtained by the LLNL model. Our model’s response is very similar to that of the LLNL model. That is, the increase in albedo reduced precipitation in the area of perturbation and slightly increased precipitation in the adjacent regions. In the perturbation area, the reduction of 13% in the zonal precipitation can be compared to 9% obtained in the LLNL model. These changes in precipitation were caused by both dynamical and hydrological factors, i.e., changes in subsidence and evapotranspiration. The model produced a cooling of the surface, also reported for the LLNL model. Thus, we see that our results agree with those of the earlier test with the LLNL model in confirming the climatic impact of albedo modification (when water availability is held constant).

In Table 5 we present the changes in mean hemispheric precipitation \( \Delta P \) and surface temperature \( \Delta T_s \) for the ice-feedback and no ice-feedback versions of the model. The results show a very slight influence of biofeedback on \( \Delta T_s \). However, there is a greater

| \( \Delta T_s \) (°C) | -0.78 | -0.82 | -1.08 | -1.15 |
| \( \Delta P \) (cm yr\(^{-1}\)) | -1.8 | -1.3 | -2.0 | -1.6 |
Influence on $\Delta \bar{P}$. As far as ice feedback is concerned, it can be seen that it merely amplifies the changes in $\bar{P}$ and $T_*$, but does not change the qualitative results. Therefore, we will now discuss only the results obtained with the no ice-feedback version.

Figure 16 shows the latitudinal distribution of the change (perturbed minus control) in precipitation only up to 60° latitude, since north of 60° the change is negligible. Here we see that in the region 20–30°N (adjacent to the perturbation) the increase in precipitation is 4 cm yr$^{-1}$ in the no-feedback run and 7 cm yr$^{-1}$ in the biofeedback run, an amplification of almost a factor of 2 due to biofeedback. Since elsewhere any differences between $\Delta P$ in these two runs were negligible, the decrease in the mean hemispheric precipitation (due to albedo perturbation) was moderated from 1.8 cm yr$^{-1}$ without biofeedback to 1.3 cm yr$^{-1}$ with biofeedback, a reduction of 25%. It should be emphasized that the above changes in sensitivity are small compared to the total rainfall, both regional and mean hemispheric. Therefore, our results on the impact of biofeedback should be considered as qualitative, regardless of the specific numbers obtained.

Let us give an interpretation of our two experiments, with and without inclusion of biofeedback, from the viewpoint of climate changes.

A perturbation of the surface albedo was imposed in the transition area between the Grassland and Semidesert zones. Since our model is stationary and has a unique steady-state solution, the albedo perturbation must be fixed for the duration of the experiment, i.e., the “perturbed” value of the albedo must be held constant until the system reaches its steady state. This can be realized in the real world through a persistent continuous destruction of vegetation (e.g., overgrazing).

There is no explicit time-dependence in this model, and so we can consider only the final equilibrium climate, i.e., the steady state. Since the time needed for the earth’s atmosphere to reach its equilibrium under prescribed external forcing would be much shorter than the relaxation time of geobotanic changes, we can interpret the two steady states obtained in the no-biofeedback and biofeedback runs as follows. When the surface state parameters are prescribed everywhere (as in our no-biofeedback run and the LLNL test), the steady state can be regarded as a simulation of the relatively rapid climatic response to the imposed fixed regional albedo change. On the other hand, when the surface state parameters are prescribed only in the perturbation area and vary everywhere else (as in the biofeedback run), the steady state can be regarded as a simulation of the long-term climatic response to a persistent regional albedo change after all geobotanic types outside the prescribed area have come into equilibrium with the climate.

In summary, our results on the model’s validation indicate that:

1) The model is able to reproduce the climate and geobotanic zones fairly close to those observed.
2) The sensitivity of the computed climate to solar constant variations agrees with that in OA.
3) The climatic response to an increase in surface albedo of the semi-arid zone is in good agreement with that simulated by the LLNL model.

In addition, preliminary tests on the effect of biofeedback indicate that:

1) The incorporation of biofeedback hardly affects the model climate’s sensitivity to solar constant variations.
2) The incorporation of biofeedback is desirable when simulating the long-term impact of a fixed surface albedo perturbation.

5. Conclusions

A climate model including an interactive land surface, based upon zonally averaged mean annual hemispheric model (Ohring and Adler, 1978), has been developed. To introduce biofeedback, a simple parameterization which makes the surface state parameters, albedo and water availability, functions of climatic variables has been suggested.

The model demonstrated its ability to reproduce the observed climate and geobotanic zones. A zonally averaged model turned out to be a suitable framework for this purpose.

The results of our experiments on sensitivity to solar constant variations suggest that the importance of the biosphere–albedo feedback mechanism may have been overestimated in previous studies. This overestimation could have been due to the representation of albedo as a function of temperature instead of through a more realistic expression using a surface type index given by a combination of thermal and moisture factors. Our study shows that the relative

![Fig. 16. Change in precipitation (perturbed minus control) for the model with (dashed line) and without (solid line) biofeedback.](image-url)
changes in these factors can be such that the net effect on the geobotanic state may become negligible. This leads us to believe that vegetation cover is relatively insensitive to small solar constant variations even though these variations do lead to temperature changes which affect living organisms. In addition, small solar constant variations cause the surface state parameters to change so little that biofeedback does not produce any noticeable difference in results from those obtained without biofeedback. Therefore, we conclude that biofeedback is probably of little importance in the simulation of changes in a zonal climate due to small perturbations in some global external parameters (such as the solar constant). Additional experiments on the doubling of atmospheric CO₂ content showed the same effect.

As an additional check of the model, a sensitivity experiment has been performed to compare the model's response to a modification of land surface albedo with that obtained by other investigators with another zonal atmospheric model. The agreement between the two has been discussed in Section 4.

An analysis of the change in precipitation in the two model runs (with and without biofeedback) suggests that sometimes it can be influenced by biofeedback. It appears desirable to have biofeedback incorporated in climate models when simulating long-term impacts of surface alterations.

We see the following prospects for the improvement and further application of the suggested approach for introducing biofeedback:

1) It is advisable to incorporate the suggested surface parameterization (Section 2, zonal case) in another zonal model containing a better description of humidity, precipitation and cloudiness (e.g., where a moisture transport equation is included).

2) The importance of changes in land biota for the carbon cycle has been emphasized in a number of studies (e.g., Bolin, 1977). A possible further extension of the suggested parameterization would be to make atmospheric CO₂ content dependent on the geobotanic state.

3) Variations in momentum exchange due to changes in the biosphere can be parameterized by making the drag coefficient a function of the geobotanic state.

4) It would be of interest to employ the suggested approach (Section 2, general case) in a simple three-dimensional model in which the annually averaged precipitation and radiation balances were available at each computational year of a long-term simulation.

In conclusion, we note that since this study was based on a very simple model, our results should be regarded as tentative and merely as an illustration of the suggested approach.

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