Energy Balance Climate Models: A Reappraisal of Ice-Albedo Feedback

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(Manuscript received 22 February 1977, in revised form 11 April 1977)

ABSTRACT

Disagreement exists, with regard to different types of climate models, concerning the influence of ice-albedo feedback upon the stability of the present global climate. In view of this we have reexamined the empirical relationship between zonal albedo and temperature for use in zonally averaged energy-balance climate models, and conclude that ice-albedo feedback constitutes a relatively mild climate feedback mechanism, amplifying global climate sensitivity by roughly 25%.

1. Introduction

Considerable emphasis has recently been placed upon modeling the terrestrial climate as a means of estimating global sensitivity to a variety of possible climatic perturbations. Schneider and Dickinson (1974) review the various feedback mechanisms which can influence climate change; they further summarize a hierarchy of models which may be employed to incorporate such mechanisms. Conventionally ice-albedo feedback is regarded as the feedback mechanism of greatest importance, and the simplest climate model capable of incorporating this feedback is the zonal annual model, which allows the zonal albedo to be temperature dependent and couples this to the global energy balance through inclusion of poleward advective heat transport.

The first such models were those of Budyko (1969) and Sellers (1969). Recently there have been numerous variations on these original models; for example, Budyko (1972), Schneider and Gal-Chen (1973), Held and Suarez (1974), Chylek and Coakley (1975), North (1975a, b), Ghil (1976), Gal-Chen and Schneider (1976), Frederiksen (1976) and Su and Hsieh (1976). While these have employed a variety of alterations to the Budyko and Sellers models, for the most part they have retained the ice-albedo parameterization of either Budyko or Sellers. Gal-Chen and Schneider (1976) do, however, suggest alternative possibilities, and they further illustrate that the predicted climate sensitivity is strongly dependent upon the chosen functional relationship between albedo and temperature.

A convenient measure of the sensitivity of the global climate, for the purpose of model comparisons, is the sensitivity parameter \( \beta = S_0 (dT_a/d\alpha) \), where \( T_a \) is the global mean surface temperature and \( S \) the solar constant, with \( S_0 = 1360 \) W m\(^{-2} \) denoting the current solar constant. A direct measure of the enhancement in global sensitivity due to ice-albedo feedback is achieved by introducing the parameter

\[
\gamma = \beta / \beta_0 - 1, \tag{1}
\]

with \( \beta_0 \) the sensitivity parameter in the absence of ice-albedo feedback. Since this feedback amplifies sensitivity, then \( \gamma > 0 \), while \( \gamma = 0 \) denotes no ice-albedo feedback.

For the original zonal annual climate models, the Budyko model yields \( \gamma = 1.58 \), whereas \( \gamma = 1.17 \) for the Sellers model. The agreement is reasonable, and these model results have led to the usual statement that ice-albedo feedback amplifies global sensitivity (i.e., amplifies \( \beta_0 \)) by several factors. More recently Wetherald and Manabe (1975), employing a threedimensional general circulation model (GCM), incorporated ice-albedo feedback into a climate sensitivity study, from which we estimate (see Section 4) that \( \gamma = 0.27 \), roughly a factor of 5 less than that of the Budyko and Sellers models. Of course, as emphasized by Wetherald and Manabe, there is no reason to conclude that the GCM result is more reliable than the results of Budyko and Sellers. But the discrepancy certainly strengthens the suggestion by Gal-Chen and Schneider (1976) that it would be appropriate to reexamine the ice-albedo parameterizations employed in energy-balance climate models, making use of recent and more extensive satellite data. The present study focuses on this point.

2. Albedo-temperature relationship

We consider first the ice-albedo parameterization employed by Sellers (1969), for which the dependence of zonal albedo \( \alpha \) upon zonal mean surface temperature \( T_a \) yields

\[
\frac{\partial \alpha}{\partial T_a} = \begin{cases} 
-C_T, & T_a < 10^\circ C \\
0, & T_a > 10^\circ C
\end{cases} \tag{2}
\]
with $C_T=0.009\,(^\circ \text{C})^{-1}$. Sellers arrived at this value for $C_T$ by comparing zonal albedos and temperatures at similar latitudes in the Northern and Southern Hemispheres. But as pointed out by Gal-Chen and Schneider (1976), the climatology of the two hemispheres differs. In particular, there are differences in cloud amount between the two hemispheres which should introduce a spurious effect into such a zonal albedo comparison.

Budyko (1969), on the other hand, considered latitudinal variations within the Northern Hemisphere. Specifically he incorporated ice-albedo feedback by employing a latitudinal albedo distribution which was a step function with a discontinuity at an average ice-boundary latitude. The corresponding albedos for the ice-covered and ice-free portions of the hemisphere were then determined by averaging the albedo as a function of latitude for the two respective regions, and it is the enhancement of the albedo for the ice-covered region, relative to that for the ice-free region, which induces ice-albedo feedback within his model. More recently Cess (1976) has shown that revising these two albedo values, employing recent satellite data, diminishes ice-albedo feedback within Budyko’s model by about a factor of 2. But there is yet another effect which could produce a further reduction. The albedo for the ice-covered portion of the hemisphere is enhanced, relative to that of the ice-free portion, not only because of ice cover, but also as a consequence of the albedo being dependent upon solar zenith angle. In particular, the albedo of clouds is greatly enhanced at high latitudes by this effect (Cess, 1976). Clearly the neglect of zenith angle dependence should lead to an overestimate of ice-albedo feedback in the Budyko model. A further complicating feature concerns the latitudinal dependence of cloud amount.

As an alternative procedure we assume that the latitudinal variation of zonal albedo may be employed to determine the dependence of zonal albedo upon surface temperature, while we eliminate zenith angle effects as well as the influence of latitudinal variations in cloud amount. Fig. 1 shows a schematic of the Northern Hemisphere which we employ for illustrative purposes, such that $\partial \alpha_s / \partial T_s$ and $\partial \alpha_c / \partial T_s$ are to be estimated from zonal annual albedo data, where $\alpha_s$ and $\alpha_c$ are the clear-sky and cloudy-sky albedos, respectively. In turn it follows that

$$\frac{\partial \alpha}{\partial T_s} = A_s(X) \frac{\partial \alpha_s}{\partial T_s} + \left[1 - A_s(X)\right] \frac{\partial \alpha_c}{\partial T_s}, \quad (3)$$

with $\partial \alpha / \partial T_s$ denoting the dependence of zonal albedo upon zonal surface temperature, which is the quantity of interest for a zonal climate model, while $A_s(X$) represents zonal cloud amount and $X = \sin(\text{latitude})$.

To evaluate $\partial \alpha_s / \partial T_s$, it is assumed that $\alpha_s = \alpha_s(T_s, \mu)$, with $\mu = \cos(\text{zenith angle})$, such that

$$\frac{\partial \alpha_s}{\partial T_s} = \frac{\partial \alpha_s}{\partial \mu} \frac{\partial \mu}{\partial T_s}, \quad (4)$$

with the second term on the right-hand side constituting a zenith angle correction. The total derivative $\partial \alpha_s / \partial T_s$ is determined from zonal annual data for $\alpha_s(X)$ (Vonder Haar and Ellis, 1975) and $T_s(X)$ (Crutcher and Meserve, 1970), while a similar procedure, employing computed values for $\mu(X)$, is used to determine $\partial \mu / \partial T_s$. To evaluate $\partial \alpha_s / \partial \mu$, we first estimated $\partial \alpha_s / \partial T_s$, where $\alpha_s$ is the surface albedo. This estimate was accomplished by dividing the Northern Hemisphere into ocean and land surfaces, further subdividing the land into various surface categories. The corresponding values of $\partial \alpha_s / \partial \mu$ for each latitude zone were then estimated from data given by Budyko (1974), Kondratyev (1969), miranova (1973) and Sellers (1965). We then employed this information into a clear-sky zonal albedo calculation which incorporates Rayleigh scattering together with the parameterizations of Lacis and Hansen (1974) for solar absorption by ozone and water vapor. We assume that $\partial \alpha_s / \partial T_s=0$ for latitudes below the 40-50°N latitude belt. This information is summarized in Table 1 and illustrates that $\partial \alpha_s / \partial T_s$ is strongly dependent on latitude and hence zonal surface temperature.

<table>
<thead>
<tr>
<th>$T_s$</th>
<th>45</th>
<th>55</th>
<th>65</th>
<th>75</th>
<th>85</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\partial \alpha_s / \partial T_s ,(^\circ \text{C})^{-1}$</td>
<td>-0.0037</td>
<td>-0.0085</td>
<td>-0.0121</td>
<td>-0.0156</td>
<td>-0.0252</td>
</tr>
<tr>
<td>$\partial \alpha_c / \partial \mu$</td>
<td>0.0101</td>
<td>0.0103</td>
<td>0.0071</td>
<td>0.0047</td>
<td>0.0035</td>
</tr>
<tr>
<td>$\partial \alpha_s / \partial \mu$</td>
<td>-0.272</td>
<td>-0.259</td>
<td>-0.252</td>
<td>-0.218</td>
<td>-0.201</td>
</tr>
<tr>
<td>$\partial \alpha_c / \partial T_s ,(^\circ \text{C})^{-1}$</td>
<td>-0.0010</td>
<td>-0.0038</td>
<td>-0.0013</td>
<td>-0.0016</td>
<td>-0.0045</td>
</tr>
<tr>
<td>$\partial \alpha_s / \partial T_s ,(^\circ \text{C})^{-1}$</td>
<td>-0.0006</td>
<td>-0.0031</td>
<td>-0.0054</td>
<td>-0.0080</td>
<td>-0.0145</td>
</tr>
</tbody>
</table>

![Fig. 1. Schematic representation of the Northern Hemisphere for the purpose of evaluating the zonal albedo-temperature relationship.](image-url)
it is assumed that $\alpha_c(\lambda, T, \mu) = \alpha_c(T, \mu)$. The temperature dependence of $\alpha_c$ is a consequence of the temperature dependence of the surface albedo, as manifested in $\alpha_c(T, \mu)$, due to multiple reflections between clouds and the surface. It should thus be possible to rephrase $\alpha_c$ as $\alpha_c = \alpha_c(\alpha, \mu)$. No attempt was made to pursue a multiple reflection calculation due to uncertainties in the numerous required input parameters. Instead, a strictly empirical approach was followed which employs the cloudy-sky albedo values given by Cess (1976): Simply employing a linear least-squares fit, correlation equations for the respective hemispheres are obtained as

$$\alpha_c(\text{NH}) = 0.641 + 0.258 \alpha_c - 0.494 \mu,$$  \hspace{1cm} (5a)

$$\alpha_c(\text{SH}) = 0.691 + 0.219 \alpha_c - 0.619 \mu.$$  \hspace{1cm} (5b)

The correlation equations agree reasonably well with one another, while the resulting rms error is 3.5%.

We attempted more complicated functional forms for $\alpha_c(\alpha, \mu)$, as suggested by multiple reflection models (Wiscombe, 1975; Schneider and Dickinson, 1976), but the above equations proved to be the most adequate. Note that Eqs. (5) uncouple the dependence of $\alpha_c$ upon surface temperature (through $\alpha_c$) and zenith angle. For example, from Eq. (5a), $\partial \alpha_c / \partial T_s = 0.26(\partial \alpha_c / \partial T_s)$, a result which incorporates solely albedo change related to surface temperature. This further indicates that clouds screen 74% of the ice-albedo effect for the cloud-covered portion of the hemisphere.

Employing Eq. (5a) together with Eq. (3), then

$$\frac{\partial \alpha_c}{\partial T_s} = [1 - 0.74 \alpha_c(X)] \frac{\partial \alpha_c}{\partial T_s},$$  \hspace{1cm} (6)

with $\partial \alpha_c / \partial T_s$, the required quantity for use in climate modeling. Taking $A_c(X)$ values from London (1957), the $\partial \alpha_c / \partial T_s$ results are listed in Table 1. Again it is emphasized that this manner of evaluating $\partial \alpha_c / \partial T_s$ accomplishes the goal of eliminating albedo enhancement at high latitudes due to the zenith angle dependence of the cloudy-sky albedo.

3. Global sensitivity

To illustrate application of the preceding albedo-temperature relationship, a zonal annual climate model is constructed with the sole intent of predicting the sensitivity of the current climate. From a zonal energy balance

$$(S/4)P(X)[1 - \alpha(X)] - F(X) = D(X),$$  \hspace{1cm} (7)

where $P(X)$ is the distribution function associated with latitudinal variations in insolation, $F(X)$ the outgoing zonal infrared flux, and $D(X)$ the divergence of the poleward advective flux. On a global basis

$$(S/4)(1 - \alpha_p) = \bar{F},$$  \hspace{1cm} (8)

with $\bar{F}$ denoting the global average value for $F(X)$, while $\alpha_p$ is the planetary albedo

$$\alpha_p = \int_0^1 \alpha(X)P(X)dX.$$  \hspace{1cm} (9)

The outgoing infrared flux may be expressed as (Cess, 1976)

$$F(X) = F_0 + F_1 T_s(X) - F_2 A_c(X),$$  \hspace{1cm} (10)

while for the divergence of the poleward advective flux, we simply follow Budyko (1969) and let

$$D(X) = v[T_s(X) - \bar{T}_s].$$  \hspace{1cm} (11)

The global sensitivity parameter follows from Eq. (8) to be

$$\beta = \frac{d\bar{T}_s}{dS} = \frac{\bar{F}}{d\bar{F}/dT_s + (S_0/4)(\partial \alpha_p/\partial T_s)},$$  \hspace{1cm} (12)

In order to evaluate this expression, note that from Eq. (9)

$$\frac{d\alpha_p}{d\bar{T}_s} = \int_0^1 d\bar{T}_s \frac{\partial \alpha}{\partial \bar{T}_s} = \frac{\alpha P(X)dX}{d\bar{T}_s},$$  \hspace{1cm} (13)

whereas employing Eqs. (10) and (11) within (7) we have

$$\frac{dT_s}{d\bar{T}_s} = \frac{v + (S_0/4\bar{F})P(X)(1 - \alpha)}{v + F_1 + (S_0/4\bar{F})(\partial \alpha/\partial T_s)},$$  \hspace{1cm} (14)

In turn, upon combining Eqs. (12)–(14), the final formulation for $\beta$ is

$$\beta = \frac{\bar{F} - (S_0/4)\int_0^1 P(X)(1 - \alpha)dX}{\int_0^1 \frac{1 + (v + F_1)[(S_0/4\bar{F})P(X)(\partial \alpha/\partial T_s)]}{dX}}.$$  \hspace{1cm} (15)

The input parameters for this expression are taken to be $F_1 = 1.6 \text{ W m}^{-2} (\circ C)^{-1}$, $v = 3.5 \text{ W m}^{-2} (\circ C)^{-1}$ and $\bar{F} = 235 \text{ W m}^{-2}$ (Cess, 1976), while $\alpha(X)$ is from Ellis and Vonder Haar (1976); all quantities pertain to the Northern Hemisphere. Ice-albedo feedback enters through $\partial \alpha / \partial T_s$. In the absence of this feedback, $\partial \alpha / \partial T_s = 0$, and Eq. (15) simply reduces to

$$\beta = \bar{F}/F_1 = 147^\circ C,$$  \hspace{1cm} (16)

whereas with the $\partial \alpha / \partial T_s$ values of Table 1, numerical quadrature yields $\beta = 184^\circ C$. These results are discussed in the following section within the context of other climate models.
Table 2. Summary of the ice-albedo sensitivity parameter $\gamma$ for several climate models. The reference to $C_T=C_T(X)$ for the present model refers to the use of $\delta a/\delta T_s$ from Table 1.

<table>
<thead>
<tr>
<th>Model</th>
<th>$\gamma = (\beta/\beta_0) - 1$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Present, $C_T=C_T(X)$</td>
<td>0.25</td>
</tr>
<tr>
<td>Present, $C_T=0.009$</td>
<td>1.68</td>
</tr>
<tr>
<td>Present, $C_T=0.004$</td>
<td>0.27</td>
</tr>
<tr>
<td>Budyko</td>
<td>1.58</td>
</tr>
<tr>
<td>Sellers, $C_T=0.009$</td>
<td>1.17</td>
</tr>
<tr>
<td>Gal-Chen and Schneider, $C_T=0.009$</td>
<td>0.98</td>
</tr>
<tr>
<td>Gal-Chen and Schneider, $C_T=0.004$</td>
<td>0.22</td>
</tr>
<tr>
<td>Wetherald and Manabe</td>
<td>0.27</td>
</tr>
</tbody>
</table>

4. Discussion of results

The ice-feedback sensitivity parameter $\gamma$ as given by Eq. (1) is summarized in Table 2 for the present model, employing both the $\delta a_s/\delta T_s$ results of Table 1 as well as Eq. (2) with $C_T=0.004$ and 0.009. Values of $\gamma$ for several other climate models are also summarized, and these were determined in the following manner.

For the Budyko (1969) model, $\beta=400^\circ\text{C}$, whereas $\beta_0=155^\circ\text{C}$ (Cess, 1976). With regard to the other climate models, $\beta$ was evaluated as $\beta=S_1(\Delta T_s/\Delta S)$, with $\Delta S$ corresponding to the change in solar constant as considered in each model. Thus from Fig. 7 of Sellers (1969) we estimate that $\beta=325^\circ\text{C}$, whereas $\beta_0=150^\circ\text{C}$ employing his infrared parameterization within Eq. (16), noting that $F_1=\delta P/\delta T_s$. The Gal-Chen and Schneider (1976) results follow directly from their Table 3 and refer to their STCV dynamical parameterization, giving $\beta=265^\circ\text{C}$ and $164^\circ\text{C}$ for $C_T=0.009$ and 0.004, respectively, while $\beta_0=134^\circ\text{C}$.

For the GCM results of Wetherald and Manabe (1975), their tabulated $\dot{T}_a$ values yield $\beta=185^\circ\text{C}$. To determine $\beta_0$, we note that $F_1=1.79$ W m$^{-2}$ from their infrared tabulation (their Table 3), such that Eq. (16) produces a sensitivity parameter of $130^\circ\text{C}$ in the absence of any albedo feedback. Their model, however, additionally includes H$_2$O-albedo feedback, which increases sensitivity by the factor 1.12 (V. Ramanathan, private communication), such that $\beta_0=146^\circ\text{C}$.

Our model, employing the empirical albedo-temperature relationship of Table 1 and designated by $C_T=C_T(X)$ within Table 2, yields an ice-albedo sensitivity which is substantially less than for the Budyko and Sellers models and, interestingly enough, agrees well with the GCM results of Wetherald and Manabe. We feel that the present ice-albedo feedback model constitutes the best state-of-the-art estimate with regard to empirical zonally-averaged climate models. Nevertheless, the model should be reexamined as more extensive satellite data become available. A particular uncertainty involves the conversion from $\delta a_s/\delta T_s$ to $\delta a_s/\delta T_a$ by means of Eq. (4), i.e., the zenith angle correction for the clear-sky albedo. But the model is not especially sensitive to this. For example, if we ignore the correction and correspondingly replace $\delta a_s/\delta T_s$ by $\delta a_s/\delta T_a$ within the model, the ice-albedo sensitivity parameter only changes from $\gamma=0.25$ to $\gamma=0.38$, this latter value still being substantially lower than the earlier estimates.

If, on the other hand, we erroneously include, as a temperature-dependent effect, the high-latitude albedo enhancement due to the zenith-angle dependence of $\alpha_0$, then our predicted value for $\gamma$ would be larger by roughly a factor of 4. This clearly illustrates the importance of uncoupling zenith angle and surface temperature dependences when empirically modeling ice-albedo feedback.

We have employed $C_T=0.009$ and 0.004, as defined by Eq. (2), within Eq. (15) solely for comparison with other climate models. Note from Table 2 that for $C_T=0.009$ the present model overestimates $\gamma$ with reference to the models of Sellers and of Gal-Chen and Schneider, this difference evidently being due to the use of different dynamical parameterizations. But the importance of dynamics will be suppressed for weaker ice-albedo feedback, since the role of dynamics in this energy-balance model is that of an interactive mechanism with ice-albedo feedback. This would suggest that, for the diminished ice-albedo feedback which we predict, global sensitivity should be fairly insensitive to the choice of a dynamical parameterization. This premise is supported by Table 2, since our $C_T=C_T(X)$ and $C_T=0.004$ values for $\gamma$ are nearly identical suggesting, for global sensitivity, that $C_T=0.004$ is an adequate synthesis, while our result and that of Gal-Chen and Schneider, again for $C_T=0.004$, are in good agreement, despite the discrepancy for greater ice-albedo feedback ($C_T=0.009$).

The preceding conclusion that $C_T=0.004$ constitutes an adequate synthesis for global sensitivity cannot, however, be extrapolated to zonal climate sensitivity. To illustrate this, we employ $dT_s/dT_a$, defined by Eq. (14), as a zonal sensitivity parameter. Results for both $C_T=C_T(X)$ and $C_T=0.004$ are summarized in Table 3 and illustrate substantial difference in $dT_s/dT_a$.

Table 3. Comparison of zonal sensitivity $dT_s/dT_a$ for $C_T=C_T(X)$ and $C_T=0.004$.

<table>
<thead>
<tr>
<th>Latitude ($^\circ\text{N}$)</th>
<th>$dT_s/dT_a$</th>
<th>$C_T=C_T(X)$</th>
<th>$C_T=0.004$</th>
</tr>
</thead>
<tbody>
<tr>
<td>85</td>
<td>1.51</td>
<td>0.88</td>
<td></td>
</tr>
<tr>
<td>75</td>
<td>1.09</td>
<td>0.91</td>
<td></td>
</tr>
<tr>
<td>65</td>
<td>1.05</td>
<td>0.97</td>
<td></td>
</tr>
<tr>
<td>55</td>
<td>1.01</td>
<td>1.07</td>
<td></td>
</tr>
<tr>
<td>45</td>
<td>0.93</td>
<td>1.18</td>
<td></td>
</tr>
<tr>
<td>35</td>
<td>0.94</td>
<td>0.94</td>
<td></td>
</tr>
<tr>
<td>25</td>
<td>0.98</td>
<td>0.98</td>
<td></td>
</tr>
<tr>
<td>15</td>
<td>1.01</td>
<td>1.01</td>
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</tr>
<tr>
<td>5</td>
<td>1.01</td>
<td>1.01</td>
<td></td>
</tr>
</tbody>
</table>
despite the agreement in global sensitivity. In all likelihood zonal sensitivity would also be dependent upon the choice of a dynamical parameterization.

A further point concerns cloud amount as a potential interactive feedback mechanism. Cess (1976) estimated zonal $\partial A_c/\partial T_e$, values from seasonal data and concluded that cloud amount was not an important feedback mechanism, either globally or zonally. He did not, however, account for possible interaction between cloud-amount feedback and ice-albedo feedback. To appraise this possibility, we have extended the present model to include cloud-amount feedback, employing Cess’ $\partial A_c/\partial T_e$ results, and again find that this feedback is insignificant, reducing $\beta$ from 184°C to 181°C.

It should be emphasized that the present study does not actually pertain to global sensitivity, since all input parameters refer to the Northern Hemisphere. We have determined $\partial a/\partial T_e$, values for the Southern Hemisphere and find them to be smaller than for the Northern Hemisphere. We have not, however, attempted to employ these results within the present model, since Eq. (11) does not fit the observations in the Southern Hemisphere.

Acknowledgment. This work was supported by the National Science Foundation under Grant ENG 72-04175.

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