JOURNAL OF GEOPHYSICAL RESEARCH, VOL. 103, NO. D6, PAGES 5929–5941, MARCH 27, 1998

On the climatic implications of volcanic cooling

Richard S. Lindzen and Constantine Giannitsis
Program in Atmospheres, Oceans, and Climate, Massachusetts Institute of Technology
Cambridge

Abstract. A simple energy balance model is used to investigate the response to a volcanic-type radiative forcing under different assumptions about the climatic sensitivity of the system. Volcanic eruptions are used as control experiments to investigate the role of the ocean-atmosphere coupling and of diffusive heat uptake by the thermocline. The effect of varying equilibrium climate sensitivity by varying the coupling of the atmosphere and ocean is examined, high sensitivity being associated with weak coupling. A model representing a coupled land-ocean system, with a reasonably realistic representation of the large-scale physics is used. It is found that systems with large equilibrium sensitivities not only respond somewhat more strongly to radiative perturbations but also return to equilibrium with much longer timescales. Based on this behavior pattern, we examine the model response to a series of volcanic eruptions following Krakatoa in 1883. Comparison between the model results and past temperature records seems to suggest that use of small sensitivity parameters is more appropriate. Despite the uncertainties associated with both the physics and the quantitative characteristics of the radiative forcing and the temperature anomalies produced by volcanic eruptions, the present study constitutes a possible test of different assumptions about the sensitivity of the climate system.

1. Introduction

The modeling of the climatic response to volcanic eruptions has been sufficiently successful to suggest that our understanding of the response of the land and sea temperatures is basically correct. Despite the simplicity of this case, it allows for testing various assumptions that may prove important for long-term climatic predictions. The initial response of the global mean temperature to a volcanic eruption depends primarily on the short timescale physics of the system. That is, the behavior for the first couple of years is relatively independent of the feedback mechanisms that influence the response to long-term climatic changes such as those due to changes in concentration of greenhouse gases. The point is simply that feedback mechanisms are excited by changes in surface temperature which are significantly delayed due to the heat capacity of the oceans. In our model simulations this is indeed the case, at least within the uncertainty arising from both the natural variability of the climatic system and the observational uncertainty in measuring a global scale quantity. As noted by *Lindzen* [1994], the gross changes in temperature following a single volcanic eruption, as predicted by models using equilibrium sensitivities to doubling CO_2 in the range between 0.25 °C and 6 °C, are observationally indistinguishable, insofar as the differences in the peak response differ by less than 0.15 °C.

However, as noted in the same paper, the long-term response presents some intriguing qualitative differences. The physical mechanisms leading to the adjustment of the system back to equilibrium after a radiative perturbation involve primarily atmospheric processes. But, as noted by *Hansen et al.* [1985], the coupling of the atmosphere with the surface is inversely proportional to the climatic sensitivity. Consequently, in more sensitive climates the oceans play an increasingly important role, and, as they involve much longer timescales, they keep the system out of equilibrium for longer time periods. *Lindzen* [1994] found that models with significant sensitivity maintained a residual cooling of the order of 1/3 of the peak response for periods longer than 20 years after the eruption, while less sensitive models had very small residual temperature perturbations. This could provide a simple test, based on past eruptions and temperature records, to differentiate between various assumptions about the sensitivity of the climate system. However, for the case of isolated eruptions, the magnitude of the remaining signal is so small that it is completely masked by climate "noise." In fact, as discussed in several papers [*Bradley,* 1988; *Robock and Mao,* 1995; *Hansen et al.*, 1997], even the depiction of the peak response to large volcanic eruptions is no trivial task. In particular, *Bradley* [1988] presents his results in terms of a standard deviation scale which highlights the problems posed by the rela-

Paper number 98JD00125.
0148-0227/98/98JD-00125$09.00

tively small signal-to-noise ratio. Both *Bradley* [1988] and *Hansen et al.* [1997] use composite responses for several volcanoes to get a clear signal.

It is possible to overcome these problems in the case of multiple consecutive volcanic eruptions. Beginning near the end of the 19th century, a series of major eruptions occurred, starting in 1883 with Krakatoa and continuing until Katmai in 1912. This series of events could allow us to check the effect of different assumptions about the sensitivity of the system on the long-term residual signal and compare them with the actual temperature records. In models with strong positive feedbacks (and hence high sensitivity), the residual signal from each eruption would add to the perturbation caused from the new volcano and thus produce not only a higher maximum response but also a larger long-term effect. In contrast to that, models representing less sensitive systems would produce isolated peaks in the temperature response, as the system would quickly return to equilibrium after each individual eruption.

Lindzen [1994] elaborates on those concepts and presents calculations based on a model that included only oceans and no land surfaces. In the present paper we investigate the behavior of the system using again a simple energy balance model, but this time including a representation of a coupled land-sea system in our set of equations. The presence of the land mass, characterized by a much smaller thermal capacity than the ocean, forces the global mean temperature to respond more strongly and quickly to the radiative forcing. However, due to the dynamical coupling of land and sea, the long-term response is mainly determined by the ocean. Thus, as our present calculations show, the bulk of the results of *Lindzen* [1994] remain valid.

2. Model Equations

In examining the system's response to volcanic forcing, we use globally averaged quantities both for the forcing function and for the temperature response. We represent our system by use of three coupled boxes, one for the land-covered regions that represent a surface area percentage of 30%, one for the ocean mixed layer, and finally use a third box to represent the thermocline and its role in communicating heat to and from the ocean surface (see Figure 1). The role of the thermocline in the energy transfer is represented by a simple diffusive process which serves as a surrogate for all ocean processes acting to carry heat from the mixed layer to the deeper layers of the ocean. The diffusive layer representing the thermocline is assumed to have a finite depth. We suppose a uniform upwelling pattern below the thermocline that acts to inhibit diffusion of heat to further depth.

Clearly, this model is not designed to represent in detail the physics involved but rather, through parametrization, to make clear the physical significance of the feedback processes involved, so as to allow us to form a relatively simple qualitative picture of the dynamics of the phenomenon. We will, in section 3, consider, in detail, the implications of various model choices. The equations defining our system are

$$C_{\text{land}} \frac{\partial \Delta T_{\text{land}}(t)}{\partial t} - \frac{\nu}{A_{\text{land}}}(\Delta T_{ml}(t) - \Delta T_{\text{land}}(t)) + \frac{B}{\text{gain}} \Delta T_{\text{land}}(t) = \Delta Q(t) \quad (1)$$

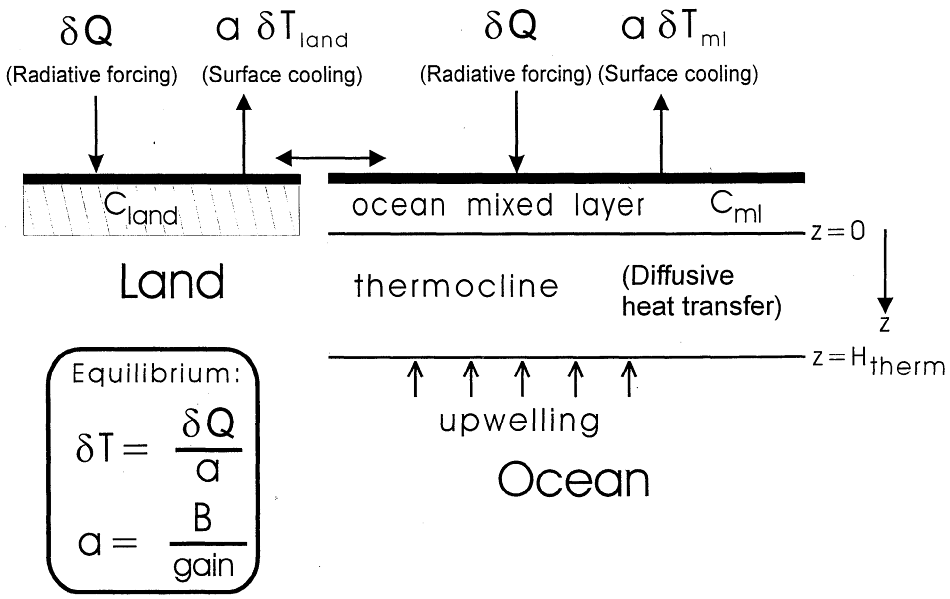

Figure 1. Geometry of simple box model for the climate system response to radiative perturbations.

$$C_{ml}\frac{\partial \Delta T_{ml}(t)}{\partial t} - \frac{\nu}{A_{\text{sea}}}(\Delta T_{\text{land}}(t) - \Delta T_{ml}(t)) + \frac{B}{\text{gain}}\Delta T_{ml}(t) + (-\lambda \frac{\partial \Delta T_{\text{therm}}}{\partial z}\bigg|_{z=0}) = \Delta Q(t) \quad (2)$$

$$\frac{\partial \Delta T_{\text{therm}}}{\partial t} = \kappa \frac{\partial^2 \Delta T_{\text{therm}}}{\partial z^2} \quad (3)$$

with boundary conditions

$$\Delta T_{\text{therm}}\bigg|_{z=0} = \Delta T_{ml} \quad (4)$$

$$\frac{\partial \Delta T_{\text{therm}}}{\partial z}\bigg|_{z=H_{\text{therm}}} = 0 \quad (5)$$

The depth of the thermocline is taken to be $H_{\text{therm}} = 400\,\text{m}$; the coefficient of eddy heat diffusivity is taken to be $\kappa = 1.5\,10^{-4}\,\text{m}^2/\text{s}$. Also the depth of the mixed layer is taken to be $H_{ml} = 75\,\text{m}$ and the thermal capacity per unit area of the land is about 30 times smaller than the heat capacity per unit area of the mixed layer. Finally, $\lambda = 620\,\text{W/m}\,^\circ\text{C}$.

The coupling between land and sea is represented by a simple linear term, proportional to the difference in averaged temperatures. The coupling coefficient is determined by tuning the system to the present climate seasonal temperature cycle. Obviously, to simulate the seasonal cycle one needs to use a model that includes a latitudinal dependence in its variables. *North and Coakley* [1979] have developed a set of equations that includes latitudinal variations and meridional heat transport by baroclinic eddies in the form of heat diffusion. Applying a perturbation to the system and taking a global average leads to the set of equations presented above, except for the absence of the thermocline in North and Coakley's model. It has been shown by *Hoffert et al.* [1980], for the case of a diffusive-advective model, that the inclusion of the thermocline does not influence significantly the response to forcings of periods smaller than 5-10 years. The same can be shown for the present model. Following the analysis of *North and Coakley* [1979], we calculated the value of ν that seems to best correspond to present atmospheric data. North and Coakley chose ν in order to simulate the zonally averaged seasonal cycle, but their choice leads to a seasonal cycle in temperature over land surfaces that is too large and does not reasonably match data. Values of ν about 5-6 times their value produce more realistic simulations (the issues are discussed in detail in the Appendix). For the present runs we chose a value of $\nu = 2.83\,\text{W/m}^2\,^\circ\text{C}$. Finally, for B we chose a value of $1.5\,\text{W/m}^2\,^\circ\text{C}$.

As to the forcing function, for the case of a single eruption, following *Hansen et al.* [1992] and *Oliver* [1976], we assume a volcanic sulfate veil being established within a period of 8 months, thus causing a linear decrease in the incoming solar radiation with time, and then decaying with an e-folding time of 13 months. For runs referring to the Mount Pinatubo eruption, we have used a maximum forcing of $4.0\,\text{W/m}^2$ [*Hansen et al.*, 1992]. For the case of the multiple consecutive volcanic eruptions, we have used a time series for the global mean radiative forcing based on Mitchell's index of stratospheric aerosol loading, covering the years between 1880 to 1920 (taken from *Robock and Free* [1995]; see also *Mitchell* [1970]). It should be noted that the uncertainty in the magnitude of the volcanic forcing for eruptions previous to Pinatubo is significant [*Robock*, 1978; *Robock and Free*, 1995; *Pollack et al.*, 1976; *Sato et al.*, 1993]. In Figure 2 one can compare different estimates of the radiative effect of aerosols resulting from the volcanoes, taken from *Mitchell* [1970], *Sato et al.* [1993], and *Robock and Free* [1995]. The details of the three curves are different; however, the agreement on the magnitude of the main eruptions is relatively good. Comparisons of results for different choices will be presented.

The parameter that is of most interest, concerning the results of the present paper, is the equilibrium climate sensitivity, given by the ratio $\Delta T_{\text{equil}}/\Delta Q$. For most general circulation models (GCMs) the value of this parameter is approximately $1\,^\circ\text{C}/(\text{W/m}^2)$; however, uncertainties about various feedback mechanisms result in large uncertainty in the sensitivity parameter. In the context of the present model the equilibrium climate sensitivity has a slightly different definition since the radiative forcing refers to radiative perturbations at the surface rather than the top of the atmosphere. Thus, for example, while the unbalanced radiative fluxes associated with a doubling of CO_2 at the top of the atmosphere take the value of $4.0\,\text{W/m}^2$, the surface fluxes are only $1.8\,\text{W/m}^2$. The equilibrium sensitivity for our model is thus given by $\Delta T_{\text{equil}}/\Delta Q_{\text{surf}} \equiv \text{gain}/B$. The "gain" parameter refers to the magnification of the equilibrium temperature response of the model to a constant external forcing with respect to the response in the absence of internal feedbacks that would change the radiative characteristics of the present atmosphere. Thus in the absence of feedback mechanisms the current model's equilibrium sensitivity is approximately $0.67\,^\circ\text{C}/(\text{W/m}^2)$ and the surface temperature response to doubling CO_2 is $1.2\,^\circ\text{C}$. In order to investigate the effect of different choices for the equilibrium climate sensitivity on the behavior of the system, we have used values of the gain ranging from 0.2 to 4.0, corresponding to equilibrium responses to doubling CO_2 in the range of $0.24\,^\circ\text{C}$ to $4.8\,^\circ\text{C}$.

It should be noted that buried in our definition of equilibrium climate sensitivity is an implicit assumption about the rigidity of the vertical temperature structure of the model's "atmosphere." In the context of a surface energy balance model the forcing used should not necessarily be the unbalanced, purely radiative flux at the surface of the three-dimensional atmosphere. The forcing at the surface should rather be some "effective" flux which depends on the dynamics of the atmospheric column above the surface, where convection and other mechanisms can act to redistribute energy between dif-

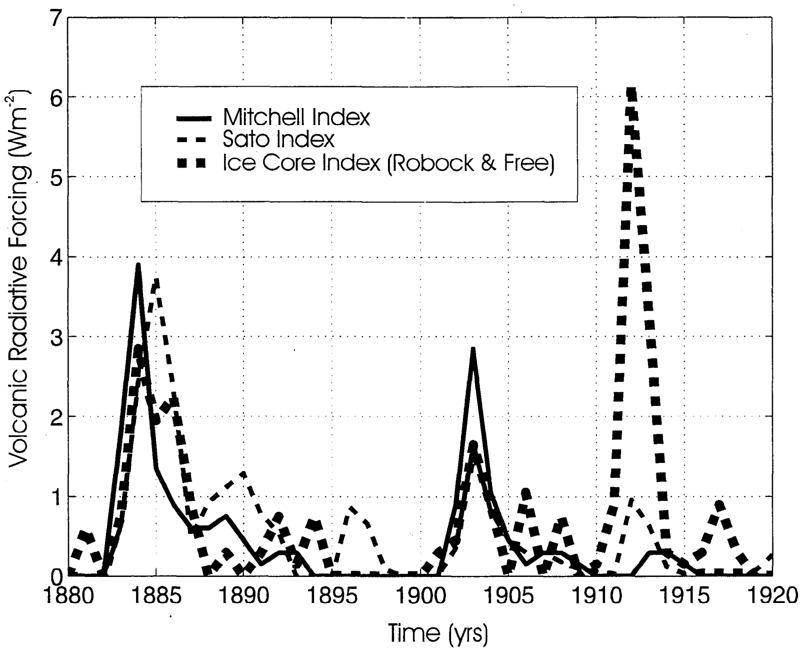

Figure 2. Comparison of different estimates of volcanic aerosol forcing time series for the period between 1880 and 1920 [*Robock and Free,* 1995; *Mitchell,* 1970; *Sato et al.,* 1993].

ferent layers. In the case of an atmosphere for which convection is assumed to act so as to keep the lapse rate fixed, the "effective" surface flux should be the flux at the tropopause, where radiative balance more nearly applies. For an atmosphere in which there is assumed to be no vertical redistribution of energy by dynamical motions, the appropriate forcing is simply the observed radiative forcing at the surface level. In principle, for differing rigidities of the vertical temperature structure, the appropriate values for the surface forcing should lie somewhere between the value of the unbalanced radiative flux at the surface and the one at the tropopause. The dependence of the "effective" surface forcing on the coupling between the surface and the interior of the atmosphere can also be shown following the methodology of *Held and Suarez* [1974], where the correspondence between the surface energy balance equations and the equations for a system including a simplified atmosphere, coupled to the surface through latent and sensible heat fluxes, is shown. Our choice of 1.8 W/m^2 for the surface forcing from doubling CO_2 implicitly assumes relatively weak atmospheric rigidity. If the lapse rate were assumed to be completely stiff and not to be affected by the heating in the interior of the troposphere due to CO_2, then the equivalent surface forcing due to double CO_2 should have been set to 4.0 W/m^2. In most GCMs the rigidity of the vertical temperature structure is greater than observed [*Sun and Held,* 1996]. Observations show significant decoupling in the vertical [*Lee and Mak,* 1994; *Sun and Held,* 1996]. However, it should be kept in mind that the above ambiguity, concerning the rigidity of the vertical temperature structure, introduces an uncertainty in our stated sensitivities, as expressed in terms of ΔT_{equil}, which may be as high as a factor of 2.

3. Comparison of Deep Ocean Parameterization With Other Models

Regarding the parameterization of heat uptake by the deep ocean, two other types of simple models have also been widely used to examine problems concerning the long-term response to external climatic forcing. The first one involves a semi-infinite, purely diffusive layer [*Hansen et al.,* 1985] and the other a finite depth diffusive-advective ocean [*Hoffert et al.,* 1980; *Harvey and Schneider,* 1985*a, b*; *Harvey,* 1986]. As discussed by *Lindzen* [1994] and *Harvey* [1986], use of a semi-infinite diffusive layer leads to very long response timescales for the system. Models including mechanisms that prohibit the diffusion of heat to great depth are considered more appropriate for answering questions related to climatic problems. In this respect the finite depth box diffusive model used here exhibits a similar behavior to the box advective-diffusive models of *Hoffert et al.* [1980] and *Harvey* [1986], for the appropriate choice of parameters. The steady state temperature structure with depth of the advective-diffusive model is characterized by a decay scale height which is given by $H = k/w$ and for the parameters used by *Hoffert et al.* [1980] is of the order of 500 m. This scale height represents roughly the depth of penetration of heat from the mixed layer. Thus, in the context of the present model, a choice of thermocline depth of the same order leads to similar behavior for timescales in the range of 10 to 100 years.

To verify the similarity of the two models, we have performed a series of calculations using various types of external forcings and have compared the results for systems of high equilibrium sensitivity, so as to maximize the importance of the deep ocean layer. The experiments include the response of the system to a step function type anomalous forcing, as well as to a linearly increasing (with time) radiative perturbation and to the two types of volcanic experiments we have described above. The results are presented in Figures 3a-3d. For reasons of simplicity we have limited the calculations to ocean-only systems. It can be shown that the results remain similar for systems that include both land and oceans. As can be seen in the figures, a choice of thermocline depth of 400 m gives quite good agreement with the advective-diffusive model, for both the suddenly applied and the linearly increasing radiative forcing. In the cases of single and multiple volcanic eruptions, one notices that the semi-infinite purely diffusive model seems to return to equilibrium quicker, relative to the finite depth thermocline model and to Hoffert's model. This is because the temperature anomalies have penetrated to greater depth during the cooling phase of the evolution and thus are less capable of influencing the behavior of the surface after the switch off of the external forcing. Leaving apart the behavior of Hansen's model, one sees again the good agreement between Hoffert's model and the model we are presently using. Therefore it is clear that the choice of the particular parameterization is not important as long as the basic physics involved are to first order correctly represented.

4. Results

The temperature response to a Pinatubo magnitude volcanic forcing, for different choices of climate sensitivity, is shown in Figures 4a-4c. The sea surface temperature (SST) signal shows a maximum perturbation in the range between $-0.15°$ and $-0.3\,°C$ for small and high sensitivities, respectively. The long-term behavior presents more apparent differences for different gain assumptions. For the case of small climate sensitivity, the SSTs quickly return to their equilibrium values after about 5 or 6 years, while for climates with high sensitivities the imbalance continues for more than two decades, and the residual signal 20 years after the eruption is as big as about one third the maximum response. Another qualitative difference in the pattern of response for different choices of sensitivity parameters is the time at which the temperature response maximizes. For low-gain systems the SST anomaly peaks 1 year after the eruption, while for high-gain systems this occurs 1 year later.

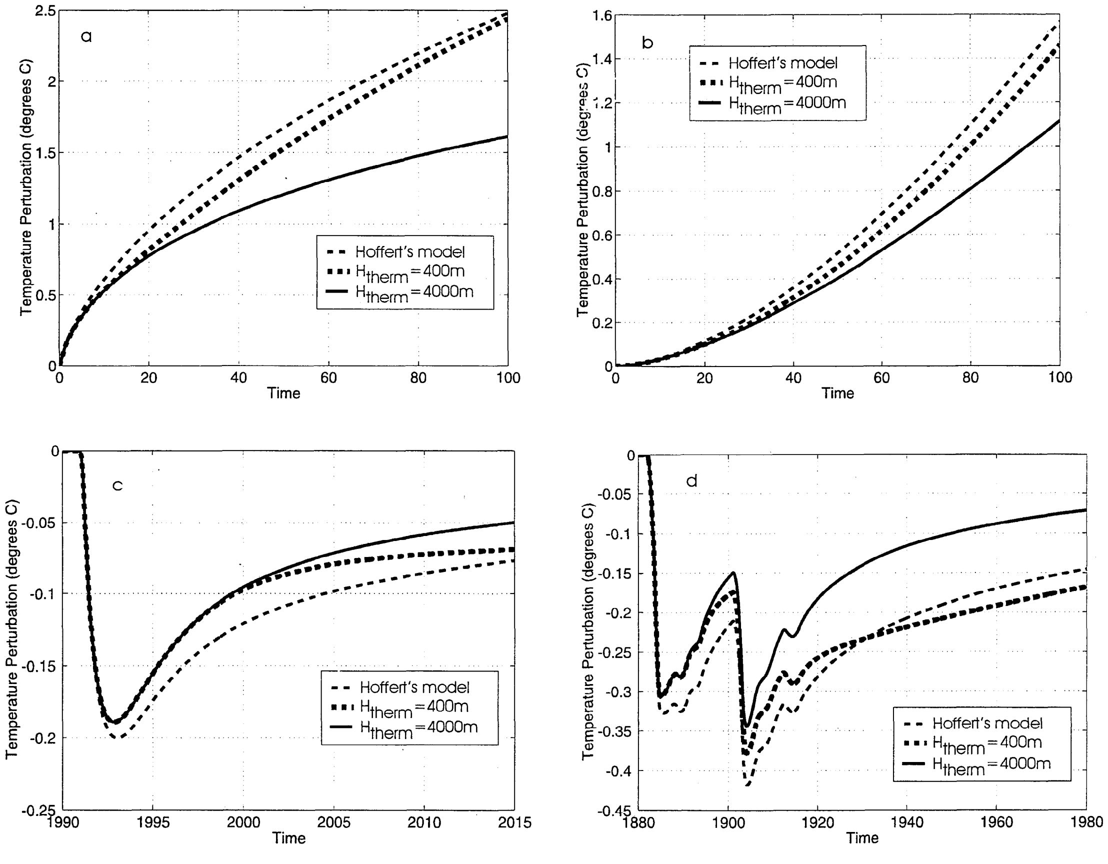

Figure 3. Comparison of the response of different models of the "deep" ocean for a system of equilibrium response to doubling CO_2 of $4.8\,°C$. The diffusive coefficient used is $k = 1.5\,cm^2/s$. The vertical velocity in *Hoffert's* model [1980] is chosen so as to keep the equilibrium temperature scale height the same as in the original version of the model. Response to (a) a step function forcing, (b) a linearly increasing forcing, (c) a single volcanic eruption, and (d) multiple consecutive eruptions.

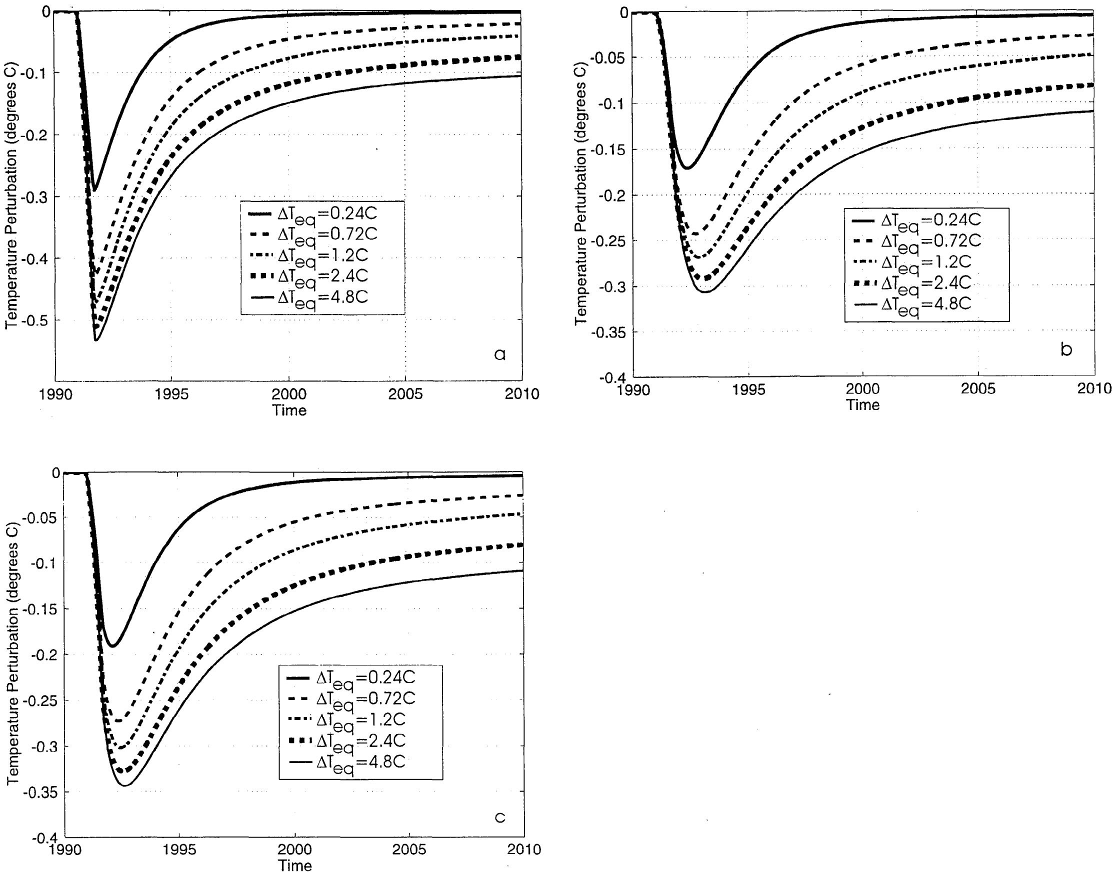

Figure 4. Response to the Pinatubo type volcanic eruption for different values of the climatic sensitivity of the system. (a) Land temperature anomalies, (b) sea temperatures, and (c) globally averaged temperature anomalies.

For land surface temperatures, the maximum response ranges between $-0.3\,^{\circ}$ and $-0.5\,^{\circ}$C and follows the maximum in the forcing with no more delay than a month. The long-term response follows the sea surface temperatures, since the two systems are closely linked through the dynamic coupling term. Thus the long-term behavior of the land temperature is controlled by the ocean diffusion mechanism. These characteristics show up in the curves for the global mean temperature. The maximum temperature perturbations for the period 1 to 2 years after the eruption lie between $-0.2\,^{\circ}$C and $-0.35\,^{\circ}$C. This is about the magnitude of the observed composite response given by *Hansen et al.* [1997] to Krakatoa, Santa Maria, Agung, El Chichon, and Pinatubo. Since at least two of these volcanoes are thought to have been about 2 times weaker than Pinatubo, our model probably slightly underestimates the response to a Pinatubo-like eruption. Thus in the context of the present model it is unlikely that we have used exaggerated volcanic forcing. On the other hand, an anonymous reviewer has suggested that according to a recent report (G. Stenchikov et al., unpublished paper, 1997), the veil from Pinatubo reached its maximum after only 5 months and that the maximum value of the forcing was $-3.2\,\mathrm{W/m^2}$. Such forcing leads to a reduction of about 25% in our calculated response to Pinatubo and an even greater underestimate of the temperature response. However, given the present degree of uncertainty, in both the forcing estimates and observations, it is unclear what the robustness of such a conclusion might be.

Concerning the results of our model, given the small separation between the different curves, compared to natural variability, it is probably not possible to distinguish between different choices of sensitivity based on the maximum temperature response. The coupling between land and sea, together with the large surface area percentage covered by sea, lead to the fact that the land surface temperature extremum is restricted to relatively low values and the global mean temperature closely follows the sea surface temperature curves. Contrary to our results, *Hansen et al.* [1992] found a maximum response for the mean global temperature of about $-0.5\,^{\circ}$C. This response magnitude seems too high. *Robock and Mao* [1995] have analyzed temperature data following the Pinatubo eruption and have estimated a globally averaged temperature anomaly maximum of about $-0.3\,^{\circ}$C. Even in the northern hemisphere where the land coverage is about 40%, the peak response they calculate is not larger than $-0.4\,^{\circ}$C. With the present model one could attain magnitudes comparable to the results of *Hansen et al.* [1992] only by choosing a much lower land-sea coupling coefficient. This, however, would lead to conflicts with observations of the seasonal temperature signal as noted above. Thus this suggests a possible inadequacy of the Goddard Institute for Space Studies (GISS) model that *Hansen et al.* [1992] are using in producing adequate coupling of the land and oceans. The difficulty of various GCMs including that at GISS in reproducing the reduced seasonality of continental interiors in reconstructions of the warm climates of the Eocene and Cretaceous may be an indication of such a problem as well [*Kerr,* 1993; *Sloan,* 1994; *Markwick,* 1994].

In this simple example of a single volcanic eruption one can see the potential for distinguishing between different assumptions for the gain, since highly sensitive climates have a strong residual signal that lasts more than two decades. However, in the case of the single volcano, neither the magnitude of the maximum response nor the magnitude of the residual signal are high enough to allow us to differentiate those cases using past temperature records, since all those signals are rendered statistically uncertain by natural variability, which for interannual or decadal timescale phenomena is typically taken to be of the order of 0.15 °C.

However, the case of the multiple consecutive eruptions allows us to get much clearer results, reflecting on the physics of the system. The results for the model runs with the multiple eruption case are presented in Figures 5a-5c. In this case one can see the large difference between systems with different climate sensitivities both in the maximum response and in the magnitude of the residual signal decades later. This difference in behavior is due to the fact that with every new eruption the temperature perturbation adds to the preexisting background temperature. Thus, since the volcanic eruptions are separated by only a few years, the residual signal has not had enough time, in the "sensitive climate" runs, to die out and the temperature anomalies keep building up. It is characteristic that in the run where the gain has been specified to be 0.2 the system returns so quickly to equilibrium that the eruptions appear as separate and the perturbation temperature time series shows only individual blips. In this case the maximum perturbation in the whole time series is not larger than −0.25 °C, just as much as an individual eruption, with a forcing of approximately the same magnitude as in the Pinatubo case, would produce. As we increase the value of the gain, the residual signal from each eruption starts becoming more important and lasting longer so the maximum signal becomes much larger and the

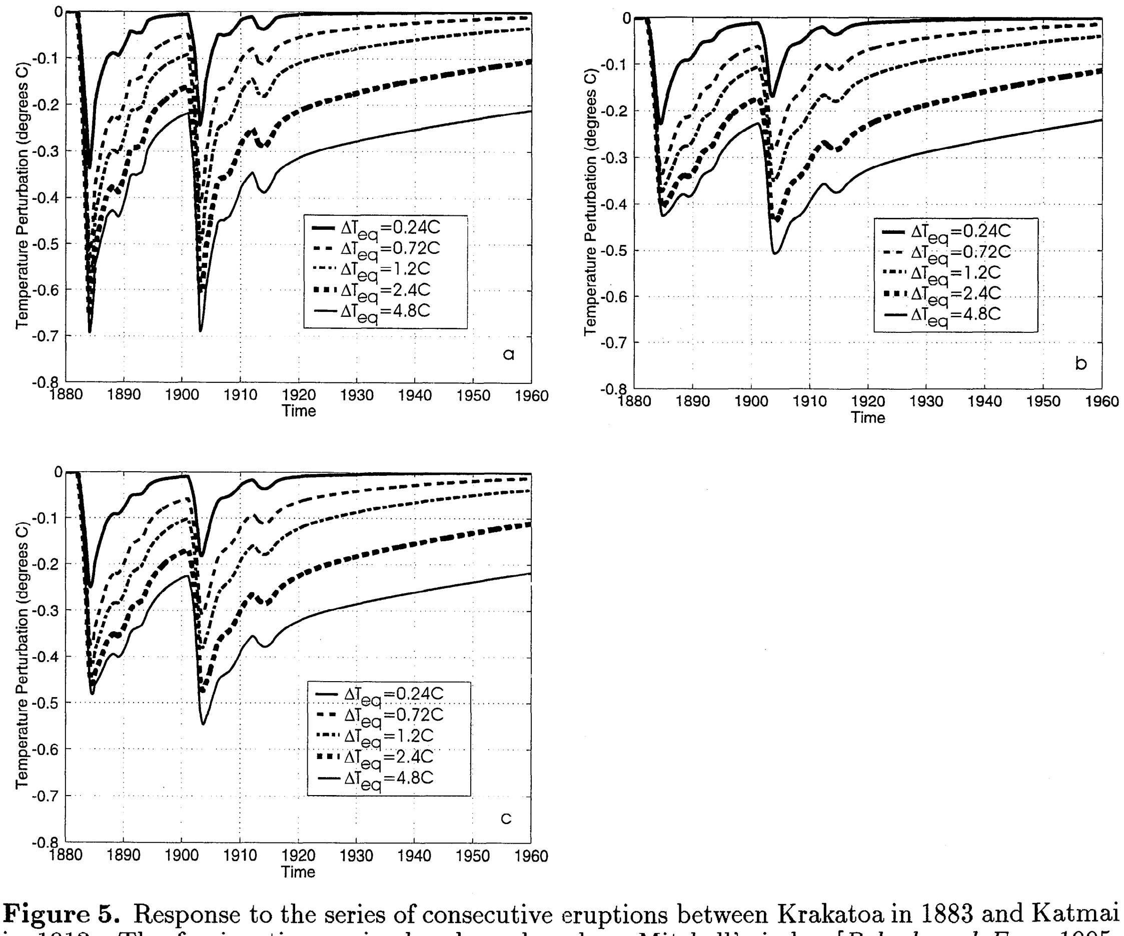

Figure 5. Response to the series of consecutive eruptions between Krakatoa in 1883 and Katmai in 1912. The forcing time series has been based on Mitchell's index [*Robock and Free,* 1995; *Mitchell,* 1970]. (a) Land temperature anomalies, (b) sea temperatures, and (c) globally averaged temperature anomalies.

dip in the temperatures is much more prominent. Thus for values of the gain parameter of the order of 2.0, the maximum global mean temperature anomaly is about $-0.45\,°C$ around 1890 and in 1905. Also in this case the dip in temperatures, following the eruptions between Krakatoa and Santa Maria, is of the order of $-0.2\,°C$ for the whole 20-year period in between those two eruptions. It should be noted that if we had chosen land-sea coupling to replicate *Hansen et al.* [1992], we would have obtained larger temperature perturbations following the various volcanos starting with Krakatoa. As far as the magnitude of the long-term residual signal is concerned, in the high-sensitivity runs it is of the order of $-0.2\,°C$ even 50 years after the last eruption. For lower values of the sensitivity parameter the signal dies out more quickly.

It is interesting that the separation of the different response curves is very small for the 2 or 3 years immediately after an eruption, while it increases as time passes. One can see either in the single volcano runs or in the multiple eruptions case that for values of the gain between 0.6 and 4.0 the temperature difference in the first few years after the initial eruption is not greater than $0.15\,°C$. Actually, there seems to be a lumping of the curves for gain greater or equal to 0.6. At the same time the curve representing a system with gain value 0.2 is separated from the rest. Therefore, if the present climate sensitivity lies anywhere in the range between 0.6 and 4.0, it is very difficult to validate any specific choice based on the short-term response of the system. However, since the separation between the curves increases as time elapses, one might be able to narrow the range of uncertainty in the magnitude of the climate sensitivity by examining the long-term behavior. The multiple eruptions case proves helpful in this respect as it increases significantly the separation between the different curves and makes it nontrivial with respect to natural variability.

Concerning the results presented above, there are some points requiring clarification. In more complicated model studies [*Robock and Mao,* 1992, 1995; *Graf et al.,* 1993] it has been found that the regional response patterns can be significantly affected by the inclusion of atmospheric dynamics. In the above papers, special patterns of behavior for high-latitude land surfaces were observed during the first 2 years after a volcanic eruption, depending on the timing of the eruption relative to the seasonal cycle. Taking into account the distribution of the observational network and the definite underrepresentation of tropical regions and of all regions in the southern hemisphere, it follows that the presence of special behaviour patterns in the northern hemisphere might mar the results when one tries to derive a global average value for the temperature perturbation. However, this does not particularly affect our results since we are more interested in testing the gross characteristics of the different responses and the long-term behavior rather than the detailed pattern of behavior of the system in the period immediately following the eruptions. Also, in using global averages for our set of equations, we have implicitly neglected any snow albedo feedback and have assumed that the radiative perturbation is not amplified at any particular latitude zone due to internal factors. In simple energy balance models that have been used in the past [*Robock,* 1978, 1984] and have been based on the Sellers and Budyko models, it was found that the snow/ice albedo feedback seriously affected the results even in a global average sense. However, there are major uncertainties in the estimates of the albedo feedback. In contrast to the results of the Budyko-Sellers type models, GCMs show much smaller albedo feedbacks, and this seems to be supported by satellite observations [*Cess et al.,* 1991]. To avoid unnecessary complexity, we choose to ignore the action of albedo feedbacks in our calculations.

As the results show, the long-term behavior is following the SST pattern due to the large thermal inertia of the ocean mixed layer and the large surface area percentage it represents. However, those results are sensitive to the influence of the thermocline, through the uptake of heat, here parameterized as a diffusion process. Obviously, GCM results including only the mixed layer cannot fully depict the long-term signal. Though the box diffusion model probably represents more fully the involved processes, the exact value of the appropriate diffusion coefficient is open to question. The present climate temperature time series requires an ocean delay time compatible with the diffusion coefficient we have used here in order to be broadly consistent with model predictions of the response to increasing greenhouse gases [*Houghton et al.,* 1992]. Use of smaller diffusivities as suggested by *Danabasoglu et al.* [1994] would lead to shorter delay times and greater incompatibilities between the past century temperature time series and model predictions [*Lindzen,* 1994]. It might be that Danabasoglu et al.'s values are not appropriate for the gross coupling averaged over all oceans.

In order to test the sensitivity of our results to uncertainties in both the volcanic forcing and the specific values of the various parameters, we performed a variety of calculations. First, to check the effect of the uncertainty in the forcing time series, we ran the multiple eruptions experiment using estimates for the aerosol forcing taken from *Sato et al.* [1993]. The results for the global mean temperature anomaly are shown in Figure 6. One can see that while the differences in the detailed structure of the temperature time series between 1883 and 1920 are significant (see Figures 5c and 6), the strength of the bulk perturbation in temperature during this whole period and of the long-term residual cooling, following the last eruption, are approximately the same as in our basic experiment. Concerning the sensitivity to changes in the various internal parameters of the model, we performed calculations using a diffusive coefficient 3 times smaller than our base case, that is, $k = 0.6\,cm^2/s$, a coupling coefficient of $1.41\,W/m^2\,°C$, and a mixed layer of depth of 30 m. The respective time series for the global mean temperatures are shown in Figure 7 for a system with an equilibrium response to doubling CO_2 of $4.8\,°C$. Again we chose a system of high sensitivity in

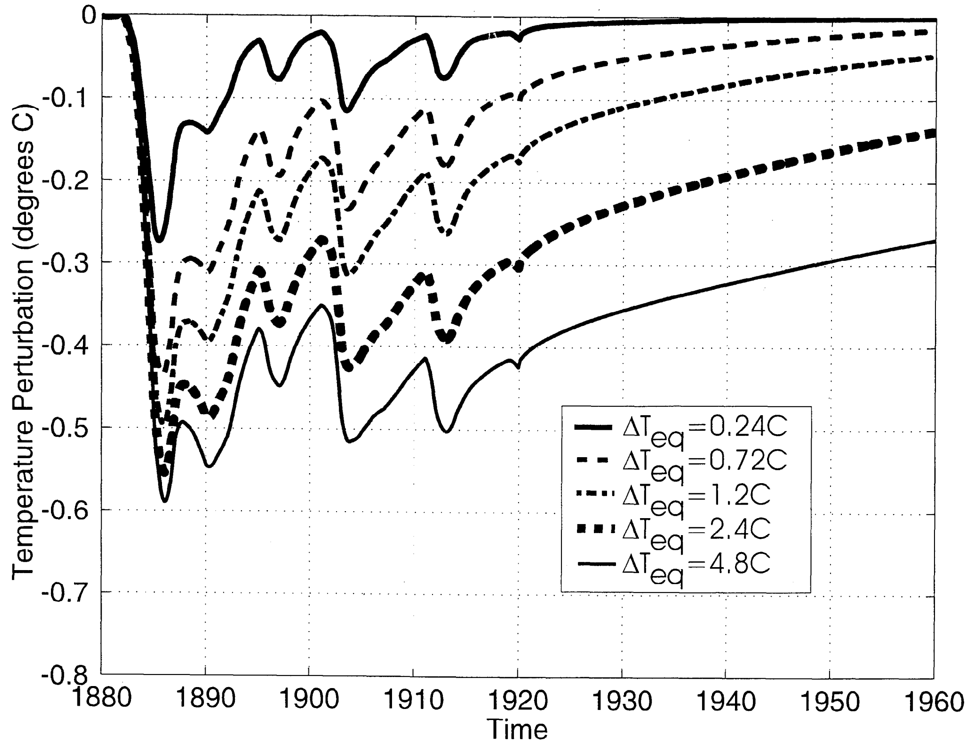

Figure 6. Globally averaged temperature response to the volcanic eruptions starting with Krakatoa, using the forcing time series of *Sato et al.* [1993].

order to maximize the importance of the deep ocean and therefore also maximize the sensitivity to any changes in internal parameters. As can be seen in this figure, regarding the gross characteristics of the behavior, the sensitivity to the choice of climatic gain is much larger than that due to the uncertainty in the mixing and coupling parameters; thus the qualitative analysis of our model results remains valid.

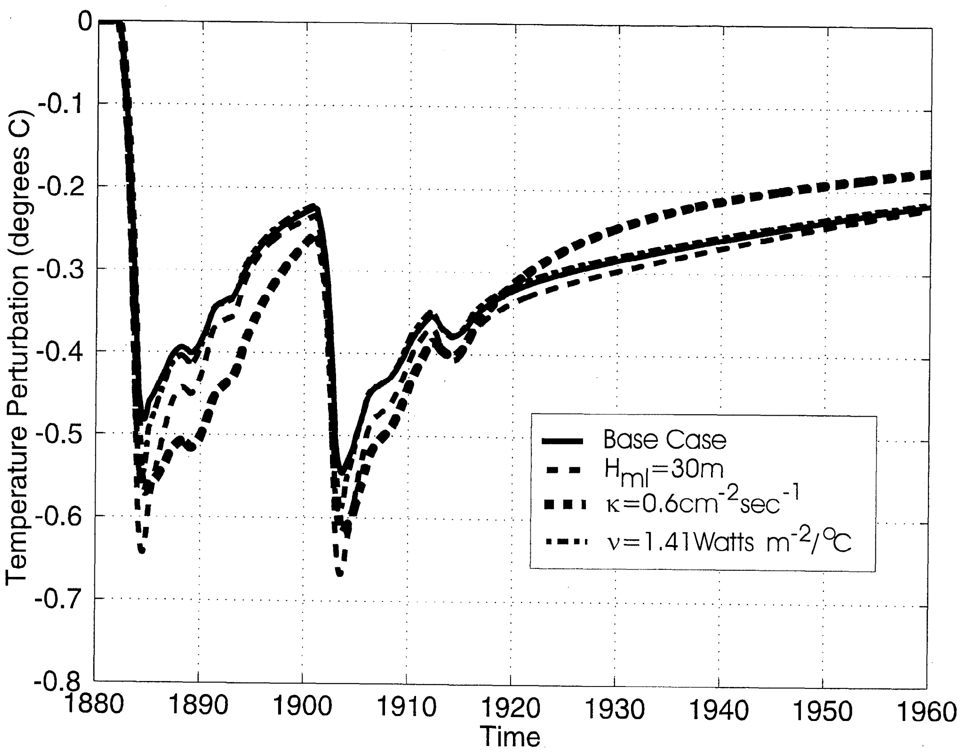

Figure 7. Sensitivity tests, for the response of a system of equilibrium response to doubling CO_2 of 4.8 °C, to changes in various internal parameters.

5. Discussion

In the work by *Lindzen* [1994] the response of a simple energy balance model to volcanic radiative forcing was examined in a system that did not include any land surfaces. The consequences of the coupling between land and sea are straightforward. Due to its large heat capacity, the ocean surface experiences only a small perturbation in temperatures. Thus coupling land and sea limits the response of the land to the forcing by carrying heat from the ocean to the land surfaces at the initial stages and thus easing the thermal forcing that is applied to them. This also forces the ocean to respond more quickly than in the sea-only model. For timescales long compared to the coupling timescale, the temperature curves of the land and sea surface follow each other closely, and it is the ocean system that mainly determines the overall behavior through the heat diffusion process. In the ocean-land system the response of the globally averaged temperature perturbation maximizes about a year after the eruption, without significant differences between systems with different sensitivities. It is not possible therefore to distinguish between different sensitivity assumptions based on the magnitude or the timing of the short-term response to a single volcanic eruption. In general, one can say that since the response during the first 2 or 3 years after the eruption is not strongly differentiated in the simple energy balance model for cases with different sensitivities, natural variability, data inadequacies and complicated dynamical feedbacks will not allow us to discriminate between various sensitivity values based on the short-term response.

On the other hand, in the case of multiple consecutive eruptions, the model results suggest that for high climate sensitivities, because of the buildup of temperature anomalies following each eruption, the dip in the temperatures is much stronger and much more prominent than in the single eruption case. Moreover, this leads to stronger long-term signals and larger separation between the predictions of models with different sensitivities. In the case of high sensitivity systems there would be a residual signal in the globally averaged temperature of about $-0.2\,°C$ for more than 30 years after the eruptions of Soufrière and Santa Maria in 1902. In order to compensate for that, one would need an additional external forcing to the system of at least as large a magnitude as the volcanic forcing. This is certainly incompatible with the estimates of the perturbation in the infrared radiation caused by CO_2 in this period. This can indeed be seen in Figure 8 where we show the results of a run that combines volcanic forcing and background forcing due to CO_2 and compare the model predictions to observed changes in the global mean temperature from 1870 to 1960 [*Hansen and Lebedeff*, 1987, 1988; *Jones et al.*, 1986]. The results for the models are normalized so that in 1880, right before the Krakatoa eruption, the predicted temperatures coincide with the "observed" global mean temperature, averaged over the period 1870 to 1880. One sees indeed the strong dips in temperatures following the eruptions of Krakatoa and Santa Maria, for the high-sensitivity case. The CO_2 forcing does not manage to reverse the sign of the temperature perturbation until about 30 years after the explosion of Santa Maria.

Of course such a comparison does not lead to ex-

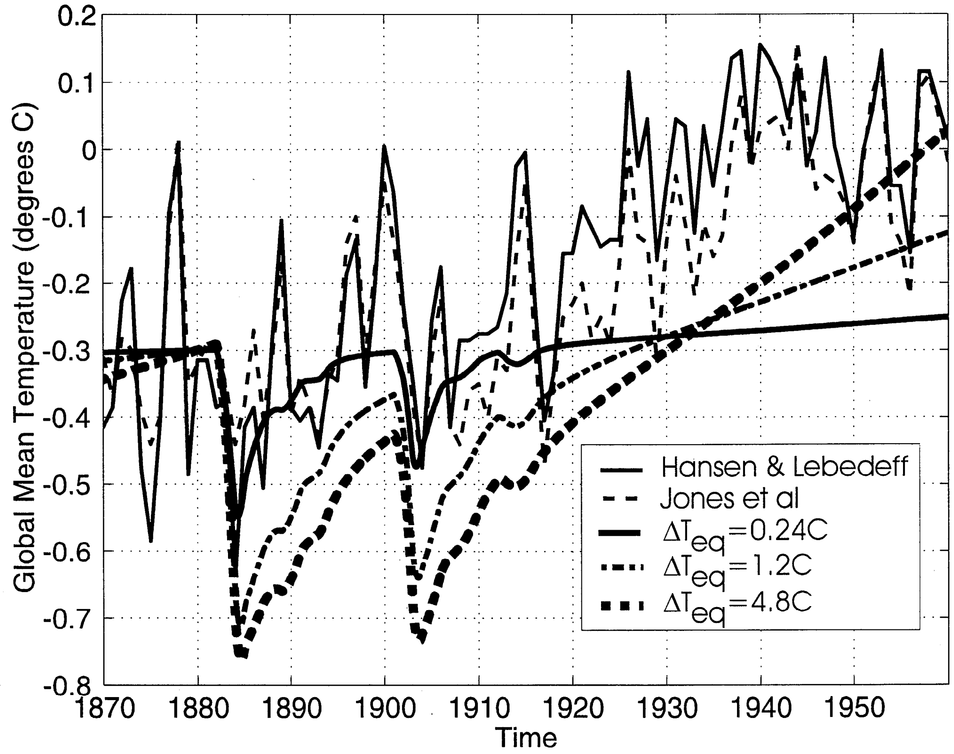

Figure 8. Comparison of observed changes in globally averaged temperature [*Hansen and Lebedeff*, 1988; *Jones et al.*, 1986] and model predictions, using combined CO_2 and volcanic radiative forcing.

plicit quantitative conclusions about possible limits on the equilibrium sensitivity of our climate system. In order to do a quantitative analysis one would have to conduct a model run that would include not only the volcanic and CO_2 forcing but also the forcing from anthropogenic tropospheric aerosols and the recorded variations in the solar irradiance, though such forcings are, themselves, uncertain. We believe that the picture would probably not change much by the inclusion of such forcings and that still the volcanic eruptions would have the largest impact in the temperature variation between 1880 and 1920. It would be interesting though to see whether a robust statistical analysis would provide limits for the equilibrium sensitivity that would allow us to narrow the range of uncertainty for this parameter and what those limits might be. Such an analysis would have to involve curve fitting the model predictions onto the observed record and would have to take into account the natural variability of the recorded temperature time series as well as the uncertainty in the prevolcano conditions, that is, the uncertainty in the "initial conditions" in 1883 when the Krakatoa eruption occurred. The parameter uncertainty of our model results and the uncertainty regarding the right choice for the CO_2 effective surface forcing would certainly affect the results of such an analysis.

In general, it is clear that the uncertainties in our results are large. For one thing, there is a large uncertainty in the magnitude of the volcanic forcing itself, although, as we have noted, it is unlikely that the forcing used here is excessive. If the magnitude of the most intense eruptions is captured correctly by any of the current estimates, then the estimate of the residual signal will be roughly correct. Apart from that, there is significant uncertainty in the estimates of the diffusivity coefficient and therefore in the timescale associated with heat uptake by the thermocline. However, as discussed by *Lindzen* [1994], in order for the predictions of high-sensitivity models to be compatible with the overall change in temperature during the last century, it would be necessary that the delay time associated with the influence of the ocean on the system be long, and for systems of equilibrium temperature change to doubling CO_2 of the order of 4°C, the appropriate delay time would have to be of the order of a century. Diffusivity coefficient values in the range between 1.0 and 2.0 cm^2/s fit better into this picture than lower diffusivity estimates. Altogether, despite the various stated uncertainties, the results of our calculations show that choices of relatively small values for the gain of the system are probably more appropriate in describing our climate system.

Appendix

North and Coakley [1979] explain in detail how it is possible by use of a simple energy balance model to simulate the seasonal temperature cycle and estimate the magnitude of the land-sea coupling coefficient that we use in our global average model. The equations they use are the following :

$$C_{\text{land}}\frac{\partial T_{\text{land}}}{\partial t} - D_o\frac{\partial}{\partial x}(1-x^2)\frac{\partial}{\partial x}T_{\text{land}} + A + BT_{\text{land}} - \frac{\nu}{A_{\text{land}}}(T_{ml} - T_{\text{land}}) = S(x,t)\,a_{\text{land}}(x,t)$$

$$C_{ml}\frac{\partial T_{ml}}{\partial t} - D_o\frac{\partial}{\partial x}(1-x^2)\frac{\partial}{\partial x}T_{ml} + A + BT_{ml} - \frac{\nu}{A_{\text{sea}}}(T_{\text{land}} - T_{ml}) = S(x,t)\,a_{\text{sea}}(x,t)$$

North and Coakley [1979] express their solutions in terms of Fourier-Legendre functions which form a natural basis for the equations above. They only retain the part of the solution that projects onto the first three Legendre polynomials and that represents the annual mean and the seasonal variations of the examined fields.

$$T_i(x,t) = T_{i,0} + T_{i,1}\cos(\omega t + \phi_i)P_1(x) + T_{i,2}P_2(x)$$

To evaluate the appropriate value for the land-sea coupling coefficient, we have used data from the National Centers for Environmental Prediction/National Center for Atmospheric Research (NCEP/NCAR) reanalysis [*Kalnay et al.,* 1996] covering the years 1982-1994. From the climatological data we have calculated the magnitude of T_1 for the land, the sea, and the zonal mean surface temperatures and compared to the prediction of the model of *North and Coakley* [1979] for a variety of choices of ν. In Table A1 we display the model prediction for the land and sea seasonal temperature variation amplitude, for both hemispheres, for a wide range of choices of the coupling coefficient. *North and Coakley* [1979] only compared the model predictions with observations for the zonal mean temperature and found good agreement for a value of ν of about 0.5 W/m^2 °C. However, this choice leads to a land temperature seasonal variation amplitude of about 40 °C for the northern hemisphere and of about 25 °C for the southern hemisphere, which in both cases are much higher than the observed values. It also leads to a sea surface temperature variation of 4 °C and 3 °C, respectively, half the observed values.

Therefore *North and Coakley* [1979] get the zonal mean temperature right by neglecting the bad agree-

Table A1. North's Model Versus Observations

	(N. Hemisphere)		(S. Hemisphere)	
ν W/m^2 °C	$T_{\text{land},1}$ °C	$T_{ml,1}$ °C	$T_{\text{land},1}$ °C	$T_{ml,1}$ °C
0.47	37.8	3.4	27.3	2.8
1.41	25.8	3.9	15.5	3.0
2.82	17.6	4.2	9.6	3.1
4.72	12.6	4.4	6.7	3.2
Observations	18.5	9.5	8.5	6.5

Predictions of *North and Coakley's* [1979] model for the amplitude of the land and sea seasonal temperature variation coefficients, for various choices of the coupling coefficient ν. Comparison with observed values calculated from NCEP/NCAR reanalysis [*Kalnay et al.,* 1996].

ment they have for both the land and sea temperature cycle. Given that the most sensitive part of the system to getting the heat transfer between land and sea right is the land surface, due to its small thermal inertia, we have chosen to use the land surface temperature as the appropriate measure of agreement between the model and observations. Clearly, to get the land surface temperature cycle right one has to choose a coupling coefficient about 6 times larger than what *North and Coakley* [1979] did.

It should be stressed that in tuning to climatology one should not seek agreement in the details, given the simplicity of the model. In that respect it is not important that the sea surface temperature variation is not tuned correctly. In order to get a sea surface temperature seasonal cycle of magnitude comparable to the observed one, one would have to use a mixed layer of only 30 m deep. However, even then the phasing of the sea surface seasonal cycle with respect to the solar forcing cycle would be very different from the observed. However, none of this is crucial in getting the magnitude of the land-sea heat exchange right, which is what we are aiming for concerning our model. Due to its parameterization, and due to the small magnitude of the mixed layer seasonal temperature variation, the heat transfer is practically only dependent on ν and $T_{\mathrm{land},1}$ and is very insensitive to variations of $T_{ml,1}$.

Acknowledgments. This work was supported by grant 914441-ATM from the National Science Foundation, and grant NAGW 525 from the National Aeronautics and Space Administration. Ten percent of this research was funded by the U.S. Department of Energy's (DOE) National Institute of Global Environmental Change (NIGEC) through the NIGEC Northeast Regional Center at Harvard University. (DOE Cooperative Agreement DEFC03-90ER61010) and through CHAMPP. Financial support does not constitute an endorsement by DOE of the views expressed in this article.

References

Bradley, R. S., The explosive volcanic eruption signal in northern hemisphere continental temperature records, *Clim. Change, 12,* 221-243, 1988.

Cess, R. D., et al., Interpretation of snow-climate feedback as produced by 17 general circulation models, *Science, 253,* 888-892, 1991.

Danabasoglu, G., J. C. McWilliams, and P. R. Gent, The role of mesoscale tracer transports in the global ocean circulation, *Science, 264,* 1123-1126, 1994.

Graf, H. F., I. Kirchner, A. Robock, and I. Schult, Pinatubo eruption winter climate effects: Model versus observations, *Clim. Dyn., 9,* 81-93, 1993.

Hansen, J., and S. Lebedeff, Global trends of measured surface air temperature, *J. Geophys. Res., 92,* 13,345-13,372, 1987.

Hansen, J., and S. Lebedeff, Global surface air temperatures: Update through 1987, *Geophys. Res. Lett., 15,* 323-326, 1988.

Hansen, J., G. Russell, A. Lacis, I. Fung, and D. Rind, Climate response times: Dependence on climate sensitivity and ocean mixing, *Science, 229,* 857-859, 1985.

Hansen, J., A. Lacis, R. Ruedy, and M. Sato, Potential climate impact of Mount Pinatubo eruption, *Geophys. Res. Lett., 19,* 215-218, 1992.

Hansen, J., M. Sato, A. Lacis, and R. Ruedy, The missing climate forcing, *Philos. Trans. R. Soc. London, 352,* 231-240, 1997.

Harvey, L. D. D., Effect of ocean mixing on the transient climate response to a CO_2 increase: Analysis of recent model results, *J. Geophys. Res., 91,* 2709-2718, 1986.

Harvey, L. D. D., and S. H. Schneider, Transient response to external forcing on $10^0 - 10^4$ year time scales, 1, Experiments with globally averaged, coupled, atmosphere and ocean energy balance models, *J. Geophys. Res., 90,* 2191-2205, 1985a.

Harvey, L. D. D., and S. H. Schneider, Transient response to external forcing on $10^0 - 10^4$ year time scales, 2, Sensitivity experiments with a seasonal, hemispherically averaged, coupled atmosphere, land, and ocean energy balance model, *J. Geophys. Res., 90,* 2207-2222, 1985b.

Held, I. M., and M. J. Suarez, Simple albedo feedback models of the icecaps, *Tellus, 26,* 613-628, 1974.

Hoffert, M. I., A. J. Callegari, and C.-T. Hsieh, The role of deep sea heat storage in the secular response to climatic forcing, *J. Geophys. Res., 85,* 6667-6679, 1980.

Houghton, J. T., B. A. Callander, and S. K. Varney (Eds.), *Climate Change 1992: The Supplementary Report to the IPCC Scientific Assessment,* 205 pp., Cambridge University Press, Cambridge, 1992.

Jones, P. D., T. M. L. Wigley, and P. B. Wright, Global temperature variations between 1861 and 1984, *Nature, 322,* 430-434, 1986.

Kalnay, E., et al., The NCEP/NCAR 40-year reanalysis project, *Bull. Am. Meteorol. Soc., 77,* 437-471, 1996.

Kerr, R. A., Fossils tell of mild winters in an ancient hothouse, *Science, 261,* 682, 1993.

Lee, W.-J., and M. Mak, Observed variability in the large-scale static stability, *J. Atmos. Sci., 51,* 2137-2144, 1994.

Lindzen, R. S., Climate dynamics and global change, *Annu. Rev. Fluid Mech., 26,* 353-378, 1994.

Markwick, P. J., "Equability," continentality, and tertiary "climate": The crocodilian perspective, *Geology, 22,* 613-616, 1994.

Mitchell, J. M., Jr., A preliminary evaluation of atmospheric pollution as a cause of the global temperature fluctuation of the past century, in *Global Effects of Environmental Pollution,* edited by S. F. Singer, pp. 139-155, D. Reidel, Norwell, Mass., 1970.

North, G. R., and J. A. Coakley, Differences between seasonal and mean annual energy balance model calculations of climate and climate sensitivity, *J. Atmos. Sci., 36,* 1189-1204, 1979.

Oliver, R. C., On the response of the hemispheric mean temperature to stratospheric dust: An empirical approach, *J. Appl. Meteorol., 15,* 933-950, 1976.

Pollack, J. B., O. B. Toon, C. Sagan, A. Summers, B. Baldwin, and W. van Camp, Volcanic explosions and climatic change: A theoretical assessment, *J. Geophys. Res., 81,* 1071-1083, 1976.

Robock, A., Internally and externally caused climate change, *J. Atmos. Sci., 35,* 1111-1122, 1978.

Robock, A., Climate model simulations of the effects of the El Chichon eruption, *Geofis. Int., 23,* 403-414, 1984.

Robock, A., and M. P. Free, Ice cores as an index of global volcanism from 1850 to the present, *J. Geophys. Res., 100,* 11,549-11,567, 1995.

Robock, A., and J. Mao, Winter warming from large volcanic eruptions, *Geophys. Res. Lett., 19,* 2405-2408, 1992.

Robock, A., and J. Mao, The volcanic signal in surface temperature observations, *J. Clim., 8,* 1086-1103, 1995.
Sato, M., J. E. Hansen, M. P. McCormick, and J. B. Pollack, Stratospheric aerosol optical depths, 1850-1990, *J. Geophys. Res., 98,* 22,987-22,994, 1993.
Sloan, L. C., Equable climates during the early Eocene: Significance of regional paleo-geography for North American climate, *Geology, 22,* 881-884, 1994.
Sun, D.-Z., and I. M. Held, A comparison of modeled and observed relationships between interannual variations of water vapor and temperature, *J. Clim., 9,* 665-675, 1996.

C. Giannitsis and R. S. Lindzen, Program in Atmospheres, Oceans, and Climate, Massachusetts Institute of Technology, 54-1711, Cambridge, MA 02139. (e-mail: giannits@mit.edu; rlindzen@mit.edu)

(Received April 16, 1997; revised December 19, 1997; accepted January 6, 1998.)