

# Reports

## Climate Response Times: Dependence on Climate Sensitivity and Ocean Mixing

**Abstract.** *The factors that determine climate response times were investigated with simple models and scaling statements. The response times are particularly sensitive to (i) the amount that the climate response is amplified by feedbacks and (ii) the representation of ocean mixing. If equilibrium climate sensitivity is 3°C or greater for a doubling of the carbon dioxide concentration, then most of the expected warming attributable to trace gases added to the atmosphere by man probably has not yet occurred. This yet to be realized warming calls into question a policy of "wait and see" regarding the issue of how to deal with increasing atmospheric carbon dioxide and other trace gases.*

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The abundances of CO<sub>2</sub> and other trace gases in the atmosphere are changing, and it is believed that this change will affect global climate (1-4). The global mean surface air warming at equilibrium ( $t \rightarrow \infty$ ) expected to result from a doubling of CO<sub>2</sub> (from, say, 300 to 600 ppm) has been estimated (1, 2) as

$$\Delta T_{eq}(2 * CO_2) = 3.0^\circ \pm 1.5^\circ C \quad (1)$$

with the range being a subjective estimate of the uncertainty based on climate modeling studies and empirical evidence for climate sensitivity.

If there were no climate feedbacks (that is, if the atmospheric temperature gradient and all other factors were fixed), the planet would have to warm by

$$\Delta T_0(2 * CO_2) = 1.2^\circ \text{ to } 1.3^\circ C \quad (2)$$

to restore the radiation balance with space after CO<sub>2</sub> is doubled (5, 6). Thus the climate sensitivity (Eq. 1) implies a net climate feedback factor

$$f = 2.4 \pm 1.2 \quad (3)$$

where  $f$  is the ratio of the equilibrium surface air warming to the warming that would have occurred in the absence of any feedbacks.

An important point in evaluating the transient response to a change in CO<sub>2</sub> is that the response time depends on  $f$ . This is illustrated by some simple but progressively more realistic systems. First, consider an atmosphereless blackbody planet ( $f = 1$ ). If the equilibrium temperature of this planet suddenly changes a small amount (say, because of a change in the solar constant), it will approach its new temperature exponentially with the blackbody e-folding time (3)

$$\tau_b = c/4\sigma T_i^3 \quad (4)$$

where  $c$  is a time-invariant heat capacity per unit of area,  $\sigma$  the Stefan-Boltzmann constant, and  $T_i$  the initial temperature.

Second, consider a planet with climate feedback factor  $f$  and fixed heat capacity  $c$ . This system has the e-folding time (6)

$$\tau = f\tau_b \quad (5)$$

because most climate feedbacks come into play only in response to the climate change (not the change in climate forcing). For example, if CO<sub>2</sub> is doubled, the initial heating is  $\sim 4 \text{ W m}^{-2}$ , independent of  $f$  or  $\Delta T_{eq}(2 * CO_2)$ . However, if positive feedbacks come into play, such as the water vapor feedback that responds to temperature, the heating decreases more gradually than in the absence of the positive feedback, and the full response is delayed.

In these examples the heat capacity  $c$  is in immediate thermal contact with the

atmosphere. This is relevant to a planet totally covered by a mixed-layer ocean, if there is negligible heat exchange between the mixed layer and the deeper ocean. For mixed-layer depth  $d_0 = 100 \text{ m}$  and  $T_i = 255 \text{ K}$ , the effective temperature of the earth,  $\tau_b$  is  $\sim 3.5$  years and thus  $f \approx 3$  yields  $\tau \approx 10$  years.

The earth is more complex than this, principally because there is significant heat exchange between the mixed layer and deeper ocean, as recognized in the CO<sub>2</sub> assessment reports (1, 2, 4) and earlier (7). Also, we must account for the fact that oceans cover only 70 percent of the earth.

The box diffusion ocean model of Oeschger *et al.* (8) provides insight into the effect of the deeper ocean on climate sensitivity. This model has a well-mixed upper layer connected to the deeper ocean by Fickian diffusion. The diffusion coefficient  $k$  is specified from observed behavior of transient tracers, such as tritium sprinkled on the ocean surface during atomic testing in the 1960's. The  $d_0$  appropriate for time scales greater than 1 year is the global mean annual maximum,  $\sim 100 \text{ m}$  (6). With this  $d_0$ , transient tracers imply an effective global  $k$  of 1 to 2 cm<sup>2</sup> sec<sup>-1</sup> (6, 8, 9).

The relation between climate response time,  $\tau$ , and  $f$  can be demonstrated by a scale analysis. Let  $d \approx (k\tau)^{1/2}$  represent depth of penetration of temperature change into the diffusive layer and  $D = d_0 + d$  represent total depth of penetration. The surface response time for the box diffusion model is proportional to the depth of penetration of the temperature change. Thus

$$\tau \approx \frac{D}{d_0} \tau_0 \approx \frac{D}{d_0} f \tau_b \quad (6)$$

where  $\tau_0$  is the (isolated) mixed-layer response time and  $\tau_b$  is the blackbody (no feedback) mixed-layer response time. In the limit  $k \rightarrow 0$ ,  $D \rightarrow d_0$  and Eq. 6 reduce to the isolated mixed-layer result (Eq. 5). For large  $k$ ,  $d \approx (k\tau)^{1/2} \gg d_0$  and

$$\tau \approx \frac{(k\tau)^{1/2}}{d_0} f \tau_b \quad (7)$$

or, solving for  $\tau$

$$\tau = kf^2 \left( \frac{\tau_b}{d_0} \right)^2 \alpha f^2 \text{ (large } k) \quad (8)$$

Typical values for  $k$  (1 cm<sup>2</sup> sec<sup>-1</sup>) and  $\tau_0$  (10 years) yield  $d \approx 170 \text{ m}$ , large enough to be nearer the diffusive regime than the isolated mixed-layer regime. More quantitative calculations, given below, indicate that for these values  $\tau$  is nearly proportional to  $f^2$  for the box diffusion model.

Before examining this model more quantitatively, we incorporate the influence of continental regions on the heat flux into the ocean surface. The effect of continental regions on climate response times was investigated previously in numerical modeling studies (10, 11). Heating over land areas can, by atmospheric transports, lead to a greater heat flux into the ocean than would occur over the same area of an all-ocean planet. The heat flux (in watts per square meter) into the ocean is proportional to the deviation of the ocean surface temperature from its equilibrium value (6):

$$F = \frac{F_0(2 * CO_2)}{\Delta T_{eq}(2 * CO_2)} (\Delta T_{eq} - \Delta T) \\ = \frac{F_0(2 * CO_2)}{f \Delta T_0(2 * CO_2)} (f \Delta T_0 - \Delta T) \quad (9)$$

where  $\Delta T$  is the ocean surface temperature departure from the 1 \* CO<sub>2</sub> equilibrium reference state,  $\Delta T_{eq}$  is the equilibrium departure for current atmospheric composition,  $F_0(2 * CO_2)$  is the flux into the ocean after CO<sub>2</sub> is doubled and stratospheric temperatures equilibrate but before the ocean temperature responds, and  $\Delta T_0$  is the equilibrium temperature change with no feedbacks ( $f = 1$ )—a measure of the radiative forcing of the climate system for a given change of atmospheric composition, independent of uncertainty in true climate sensitivity.

The proportionality constant in Eq. 9 can be obtained from a 2 \* CO<sub>2</sub> experiment in a three-dimensional (3-D) model with realistic geography. Our 3-D model yields  $F_0(2 * CO_2) = 4.3 \text{ W m}^{-2}$  (5). On the basis of one-dimensional (1-D) calculations, we estimate that the flux into the surface of an all-ocean planet with this radiation scheme would be 3.5 to 4.0 W m<sup>-2</sup>. Thus, in this model, heating of continental regions increases the flux into the ocean (per unit of ocean area) by 10 to 20 percent.

The radiative forcing  $\Delta T_0$  as a function of CO<sub>2</sub> concentration  $\chi$  (parts per million), computed with the radiation scheme of our 3-D model (6), is fitted by  $\Delta T_0(\chi) = \ln [1 + 1.2\chi + 0.005\chi^2 + 1.4 \times 10^{-6}\chi^3]$  (10)

to better than 1 percent for  $\chi$  between 100 and 1000 ppm. The absolute accuracy of Eq. 10 is estimated to be 10 to 15 percent as a result of uncertainties in CO<sub>2</sub> absorption coefficients and numerical approximations in the modeling (12).

Equations 9 and 10 and the box diffusion model for ocean heat storage allow a numerical solution for the temperature

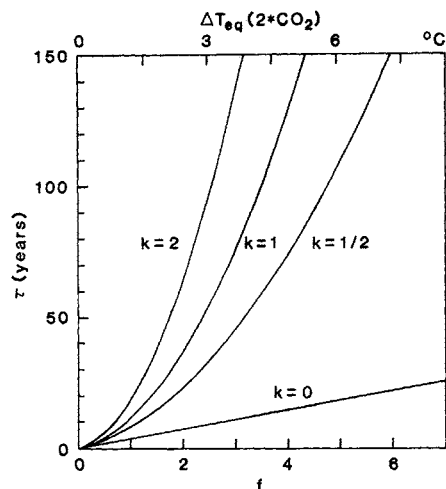


Fig. 1. Ocean surface response time (time to reach  $1 - e^{-1}$  of equilibrium response) for the 1-D box diffusion ocean model as a function of climate feedback factor,  $f$ , or climate sensitivity to doubled CO<sub>2</sub>,  $\Delta T_{eq}(2 * CO_2)$  ( $k$  is the vertical diffusion coefficient in square centimeters per second; the mixed-layer depth  $d_0 = 100 \text{ m}$ ). Results depend strongly on  $k$ , as illustrated, but only slightly on  $d_0$  for the cases in which  $k \geq 0.5 \text{ cm}^2 \text{ sec}^{-1}$ . Calculations are based on Eqs. 9 and 10, with an instant doubling of CO<sub>2</sub> from 300 to 600 ppm.

trend at the ocean surface for any CO<sub>2</sub> scenario. Let us first consider the idealized case of instant doubling of CO<sub>2</sub> from 300 to 600 ppm. The resulting time required for the mixed layer to reach 63 percent of its equilibrium response,  $\tau$ , is shown in Fig. 1 as a function of  $f$  or  $\Delta T_{eq}(2 * CO_2)$ .  $\tau$  is 15 to 25 years for climate sensitivity 1.5°C but 50 to 100 years for climate sensitivity 3°C, if the diffusion coefficient derived from ocean tracers ( $k = 1$  to  $2 \text{ cm}^2 \text{ sec}^{-1}$ ) is used.

We next let CO<sub>2</sub> increase linearly from

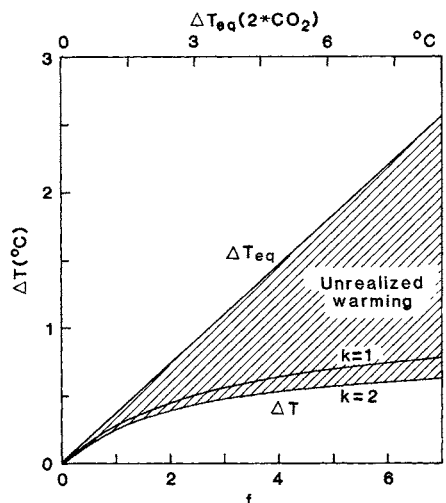


Fig. 2. Ocean surface warming ( $\Delta T$ ) and the equilibrium warming ( $\Delta T_{eq}$ ) due to CO<sub>2</sub> added to the atmosphere in the period 1850 to 1980 for the 1-D box diffusion ocean model as a function of  $f$  or  $\Delta T_{eq}(2 * CO_2)$ .

270 ppm in 1850 to 315 ppm in 1958, and thereafter as observed by Keeling *et al.* (13). The results (Fig. 2) show that a large part of the equilibrium CO<sub>2</sub> warming is not yet realized, the unrealized fraction depending strongly on  $f$  or  $\Delta T_{eq}(2 * CO_2)$ . If climate sensitivity is 3°C or greater for doubled CO<sub>2</sub>, most of the expected equilibrium warming due to the CO<sub>2</sub> increase since 1850 probably has not yet occurred; this must be all the more true for other trace gases, whose greenhouse effect is dominated by chlorofluorocarbons added since 1960 (14). This yet to be realized warming calls into question a policy of "wait and see" and "if necessary, make mid-course corrections" regarding the issue of how to deal with increasing atmospheric CO<sub>2</sub> and other trace gases.

The Carbon Dioxide Assessment Committee (4) attempted to infer climate sensitivity from the Northern Hemisphere surface air warming of 0.5° to 0.6°C in the period 1850 to 1980 by assuming that the warming was due to the increase in CO<sub>2</sub> during that period. With the added assumption that the temperature trends are similar for the ocean surface and hemispheric surface air, they concluded that this evidence indicates a climate sensitivity in the lower part of the 1.5° to 4.5°C range estimated by Charney (1) and Smagorinsky (2). However, their analysis assumed  $\tau = 15$  years, independent of  $f$ . As shown above,  $\tau$  depends strongly, at least linearly, on  $f$ . If this dependence is included, along with uncertainties in the actual warming, the 1850 CO<sub>2</sub> abundance, and other variable climate forcings, it is difficult to set an upper limit on climate sensitivity by this method. As shown in Fig. 2, a small uncertainty in  $\Delta T$  leads to a large uncertainty in  $f$  or  $\Delta T_{eq}$ .

The box diffusion model is a gross simplification of vertical transport in the ocean, and the proportionality of  $\tau$  to  $f^2$  depends on this diffusive representation as well as on the assumption that a small heat perturbation mixes as a passive tracer. If one used a two-box ocean model with a mixed layer connected to a well-mixed thermocline by a fixed exchange rate,  $\tau$  would vary linearly with  $f$ . However, such a two-box model is less realistic than the box diffusion model, failing to represent the increasing penetration of transient tracers with time. Because their observed penetration does increase with time,  $\tau$  must increase more strongly than linearly with  $f$  for the real ocean. A more precise conclusion depends on obtaining better data on the world ocean circulation, especially the vertical mixing processes.

The assumption that a heat perturbation mixes as a passive tracer may break down as the climatic warming increases. In the ocean model of Bryan *et al.* (15), a warm anomaly of 0.5°C penetrates significantly (~25 percent) less than a similar cold anomaly. Furthermore, global warming will be accompanied by changes in evaporation, precipitation, and wind stress over the ocean surface, and possibly by the addition of fresh water from melting ice sheets—all of which may affect the rate of ocean mixing. There is evidence that some mechanisms of ocean overturning are capable of sudden changes (16), and the paleoclimate record reveals cases of large warming within periods of no more than several decades (16, 17). Thus we cannot exclude the possibility that the climate may at some point undergo a rapid transition to the equilibrium climate for current atmospheric composition.

The existence of unrealized warming complicates the CO<sub>2</sub> and trace gas issue and limits the near-term effectiveness of reductions in greenhouse gas emissions. The strong dependence of this unrealized warming on the equilibrium climate sensitivity emphasizes the importance of narrowing uncertainties about the strength of climate feedback processes. This will require better understanding of many components of the climate system including clouds, the cryosphere, biogeochemical cycles, ocean mixing, vegetation, and the land surface.

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- Our 3-D model [*J. Atmos. Sci.* **111**, 609 (1983)] averages gaseous absorption over broad ranges for computational efficiency. This radiation scheme yields a forcing  $\Delta T_0 = 1.2^\circ\text{C}$  for doubled CO<sub>2</sub> (300 to 600 ppm) and an initial flux  $F_0 = 4.3 \text{ W m}^{-2}$  into the ocean in the 3-D model. A more detailed radiation scheme with fine spectral resolution yields  $\Delta T_0 = 1.26^\circ\text{C}$  for 1-D calculations with a global mean atmospheric profile and would be expected to increase  $F_0$  proportionately, without affecting  $\tau$ . The latter radiation scheme has been compared with a wide range of other models, including line-by-line calculations,

- with results agreeing within ~10 percent [F. M. Luther and Y. Fouquart, *WMO Rep. WCP-93* (1984)]. Other published values for  $\Delta T_0$  are 1.33°C (4) and 1.3°C [V. Ramanathan, *J. Atmos. Sci.* **38**, 918 (1981)].
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## Acid Deposition, Smelter Emissions, and the Linearity Issue in the Western United States

**Abstract.** *The variation in sulfur dioxide emissions from nonferrous metal smelters in the western United States over a 4-year period is compared with the variation in sulfate concentrations in precipitation in the Rocky Mountain states. The data support a linear relation between emissions and sulfate concentration. The geographic separation of emissions sources and precipitation monitors indicates a sulfur transport scale exceeding 1000 kilometers.*

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The relation between SO<sub>2</sub> emissions at particular sources and acid deposition at distant receptors has been the subject of several investigations (1, 2). Because of the sparsity of historical deposition data, the lack of large excursions in emissions

in recent years, and the high source density with low gradients, empirical studies of SO<sub>2</sub> emissions and deposition in eastern North America have not permitted the determination of definitive spatial or temporal relations between emissions at sources and deposition at distant receptors (2).

The intermountain region of the western United States (from the Sierra crest to the continental divide) is an especially suitable region for studying long-range transport and source-receptor relations because it is characterized by a few large

Table 1. Annual VWM sulfate concentrations at NADP recording stations (Fig. 1). All precipitation-monitoring stations in the intermountain region downwind (9) of the smelters that had data for at least 30 weeks during 10 months in 1981 (when a large emissions increase occurred) and 1982, including three stations just east of the continental divide, are listed. The error due to imprecision in sampling and chemical analysis was less than 10 percent (18). Quality control of sampling and analysis procedures by monitoring stations and the central analytic laboratory has been described (19). VWM sulfate concentration was averaged over all stations for each year to yield a CVWM; the difference between CVWM concentrations for 1981 and 1982 was statistically significant ( $P < 0.01$ ,  $t$ -test) and greater than 0.47 mg per liter ( $P = 0.05$ ,  $t$ -test). Values in parentheses for 1982 represent the standard deviation in monthly mean sulfate concentration, which is a measure of site-specific within-year variation. Data for the Hubbard Brook and Hopland sites, provided for comparison, are not included in the CVWM.

Site (number)	Annual mean sulfate concentration (milligrams per liter)			
	1980	1981	1982	1983
Alamosa, Colorado (1)		1.81	1.48 (0.75)	1.39
Sand Spring, Colorado (2)	1.05	1.60	0.96 (0.27)	1.05
Manitou, Colorado (3)		1.62	0.98 (0.42)	0.91
Pawnee, Colorado (4)	1.50	2.17	0.83 (0.49)	1.14
Rocky Mount, Colorado (5)		1.57	0.91 (0.35)	0.83
Yellowstone, Wyoming (6)		1.56	0.79 (0.27)	0.53
Craters of the Moon, Idaho (7)		0.91	0.61 (0.29)	0.65
Organ Pipe, Arizona (8)		2.14	0.87 (0.30)	0.64
CVWM (5)	1.28	1.60	0.86	0.85
Standard error (6)	0.22	0.14	0.06	0.08
Hubbard Brook, New Hampshire	2.45	2.25	2.13	1.65
Hopland, California	0.50	0.33	0.34	0.33