

## On the feedback of stratospheric clouds on polar climate

Daniel B. Kirk-Davidoff, Daniel P. Schrag, and James G. Anderson

Department of Earth and Planetary Sciences, Harvard University, Cambridge, Massachusetts, USA

Received 2 January 2002; revised 20 March 2002; accepted 1 May 2002; published 15 June 2002.

[1] Past climates, such as the Eocene (55 - 38 Ma), experienced dramatically warmer polar winters. Global climate models run with Eocene-like boundary conditions have under-predicted polar temperatures, a discrepancy which has stimulated a recent hypothesis that polar stratospheric clouds may have been important. We propose that such clouds form in response to higher CO<sub>2</sub> via changes in stratospheric circulation and water content. We show that the absence of this mechanism from models of Eocene climate may be attributable to poor vertical resolution in the neighborhood of the tropical tropopause. This may cause the models to underestimate future greenhouse warming. *INDEX TERMS*: 1620 Global Change: Climate dynamics (3309); 3362 Meteorology and Atmospheric Dynamics: Stratosphere/troposphere interactions; 3319 Meteorology and Atmospheric Dynamics: General circulation; 3344 Meteorology and Atmospheric Dynamics: Paleoclimatology; 0340 Atmospheric Composition and Structure: Middle atmosphere—composition and chemistry

### 1. Introduction

[2] Paleoclimate data from the Eocene indicate that deep ocean temperatures were approximately 10 K warmer than at present [Douglas and Woodruff, 1981], as were polar sea surface temperatures [Zachos *et al.*, 1994]. Northern hemisphere continental interiors were warm enough throughout the year to sustain palm trees and other tropical flora and fauna, as were far northern land regions such as Ellesmere Island. Continental glaciation was essentially absent [Sloan and Barron, 1992]. By contrast, tropical temperatures appear to have been similar to present values [Schrag, 1999], or possibly somewhat warmer [Pearson *et al.*, 2001].

[3] Several studies have sought to identify the mechanisms responsible for the Eocene climate and other warm climates, such as the Cretaceous. Possible mechanisms include enhanced meridional heat fluxes due to ocean circulation changes [Toggweiler and Bjornsson, 2000] or to reorganization of atmospheric circulation [Lindzen and Pan, 1994; Farrell, 1990], enhanced greenhouse warming due to high carbon dioxide concentrations [Sloan and Barron, 1992], and reductions in global topography [Bush and Philander, 1997]. Although these modeling efforts have produced relatively warm polar temperatures, they have not been able to reproduce the above-freezing temperatures of the reconstructed Eocene wintertime polar and continental interior regions. This failure is important, because it suggests that the GCMs have neglected a strong warming mechanism, operating in warm climates, which acts preferentially in cold regions. Such a mechanism might also be operative in a future climate, warmed by increased greenhouse gas concentrations, leading to underestimates of future polar warmth by GCMs.

[4] Recently, Sloan and Pollard [1998] imposed optically thick (optical depth unity) polar stratospheric clouds (PSCs) in a GCM, in combination with increased greenhouse gas concentration, producing a temperature distribution more consistent with observations from the Eocene. The addition of PSCs to the polar night stratosphere resulted in wintertime surface warming ranging from 4 K in midlatitude regions to 20 K in some locations near the poles. They suggested that these thick PSCs might result from an increase in stratospheric water vapor due to higher fluxes of methane into the stratosphere, perhaps due to increased swamp and wetland area in a warmer world. However, while Eocene polar warmth appears to have been uninterrupted for over ten million years [Zachos *et al.*, 1994], the lifetime of methane in the modern atmosphere is quite short ( $\approx 7$  years) [Lelieveld *et al.*, 1998]. Thus, if polar warmth were dependent on high methane levels, the latter must have been maintained by consistently elevated methane production over this period. While there is isotopic evidence for large releases of atmospheric methane to the ocean and atmosphere during the late Paleocene, evidence of sustained high methane levels is lacking.

[5] While the attribution of polar warmth to high methane levels remains plausible, we believe that a direct link to CO<sub>2</sub> would be more parsimonious, because the large reserve of carbonate in the ocean gives atmospheric CO<sub>2</sub> a long lifetime, and because the association of global warmth with polar warmth is strong throughout the geological record. In this paper we propose an alternative mechanism for PSC formation as a positive feedback on the meridional temperature gradient, and on greenhouse gas concentration. Our mechanism works as follows. GCMs respond to increased greenhouse gas (GHG) concentrations by reducing the equator-to-pole temperature difference (EPTD). This result is due to the positive ice- and snow-albedo feedback in polar regions, which tends to increase the surface warming initiated by GHG increases in the polar regions. A reduction in the EPTD might also have been forced by an increase in oceanic or atmospheric heat transports due to changes in the positions and shapes of the continents [Toggweiler and Bjornsson, 2000; Lindzen and Pan, 1994].

[6] Once the EPTD were substantially reduced from its present value, we posit that the energy of gravity and planetary waves propagating from the troposphere into the stratosphere would be reduced. Evidence to support this supposition comes from several sources. A smaller surface temperature gradient from the equator to the poles implies a reduced effective potential vorticity gradient at the surface, and smaller growth rate for atmospheric eddies [Lindzen and Farrell, 1980]. GCM results for climates with reduced EPTD [Rind, 1998], and for doubled CO<sub>2</sub> [Shindell *et al.*, 1998], show such a reduction in planetary wave activity, and observations of recent trends in wave propagation into the stratosphere show a decreasing trend, along with a decrease in stratospheric overturning, as the surface climate has warmed [Fusco and Salby, 1999]. These results alone do not demonstrate a direct link between surface temperature gradient and deposition of wave pseudomomentum in the stratosphere, since a reduced meridional temperature gradient might be expected to result, via thermal wind balance, in reduced mean jet speeds, which would tend to allow a

**Table 1.** Model results for Modern,  $2 \times \text{CO}_2$ , and Eocene Simulations

Climate	$T_{eq}$ (K)	$T_p$ (K)	$\Delta T_s$ (K)	$q$ (ppmv)	$\tau$	$F_p$ $\text{W m}^{-2}$	$\Delta \text{OLR}$ $\text{W m}^{-2}$
Modern	300	255	20	6.6	0.075	1.4	-1.1
$2 \times \text{CO}_2$	303	263	20	13	0.13	2.55	-2.7
Eocene	300	285	10	17	0.22	15.3	-7.6
Eocene (SP)	300	285	10	17	0.50	36.2	-15.4

Results are 5-year averages for regions poleward of  $84^\circ$  in each hemisphere, except for stratospheric water vapor ( $q$ ), which represents the 5-year average in the lower tropical stratosphere (i.e. without any contribution from methane oxidation). Eocene (SP) shows the results for PSCs of optical depth 1 inserted at 60 mb, whenever a location is in darkness for a whole month, as in *Sloan and Pollard* [1998].

broader spectral range of planetary waves to propagate into the stratosphere [*Charney and Drazin*, 1961]. However, GCM simulations of warmer climates have generally not shown a reduction in jet speeds, because reduced meridional temperature gradients at the surface have not, in the models, been accompanied by reduced gradients aloft [*Shindell et al.*, 1998; *Bush and Philander*, 1997; *Rind et al.*, 2001].

[7] Assuming that a reduced EPTD would reduce the upward propagation of waves into the stratosphere, it should also reduce the momentum deposited there by those waves. It is this momentum deposition that drives the overturning circulation of the stratosphere [*Haynes et al.*, 1991]; the overturning acts to cool the tropical stratosphere via forced ascent and warm the polar stratosphere via forced descent [*Eluszkiewicz et al.*, 1996]. So we expect reduced wave momentum deposition to warm the tropical stratosphere and cool the polar stratosphere. Such an effect has been observed in a GCM study of the Paleocene stratosphere [*Rind et al.*, 2001], though this result is sensitive to the details of stratospheric temperature change. Warming of the tropical stratosphere, assuming it extends to the tropopause, will tend to increase the concentration of water vapor entering the stratosphere [*Weinstock et al.*, 2001] (though other factors, including aerosol burden [*Sherwood*, 2001], may influence this concentration). An increase in stratospheric water vapor would lead to more frequent, optically thicker PSCs, an effect enhanced by the cooling of the polar stratosphere [*Kirk-Davidoff et al.*, 1999]. A reduced stratospheric overturning circulation would also tend to reduce mid-latitude and polar ozone concentrations.

[8] We can express this feedback mechanism schematically as: Change in PSC Heating = [Change in PSCs due to Water Vapor]  $\times$  [Change in Water Vapor due to Changes in Climate and Circulation]  $1 \text{ cm} + 0.6 \text{ cm}$  [Change in PSCs due to change in Stratospheric Temperature]  $\times$  [Change in Stratospheric Temperature due to Changes in Climate and Circulation] or as a chain of partial derivatives:

$$dF_p = \left\{ \frac{\partial F_p}{\partial \tau} \left[ \frac{\partial \tau}{\partial q} \frac{\partial q}{\partial T_t} \left( \frac{\partial T_t}{\partial \omega} \frac{\partial \omega}{\partial \text{CO}_2} + \frac{\partial T_t}{\partial T_{eq}} \frac{\partial T_{eq}}{\partial \text{CO}_2} \right) + \frac{\partial \tau}{\partial T_{st}} \left( \frac{\partial T_{st}}{\partial \omega} \frac{\partial \omega}{\partial \text{CO}_2} + \frac{\partial T_{st}}{\partial \text{CO}_2} \right) \right] \right\} d\text{CO}_2, \quad (1)$$

where

$$\frac{\partial \omega}{\partial \text{CO}_2} = \frac{\partial \omega}{\partial M} \left( \frac{\partial M}{\partial A} \frac{\partial A}{\partial (T_{eq} - T_p)} \frac{\partial (T_{eq} - T_p)}{\partial \text{CO}_2} + \frac{\partial M}{\partial J} \frac{\partial J}{\partial \text{CO}_2} \right), \quad (2)$$

$F_p$  is the surface heating due to PSCs,  $\tau$  is the PSC optical depth,  $q$  is the polar stratospheric water vapor mixing ratio,  $T_t$  is the mean temperature of the tropical tropopause,  $\omega$  is the rate of stratospheric overturning,  $\text{CO}_2$  is the concentration of carbon dioxide,  $T_{st}$  is the

mean temperature of the polar vortex,  $M$  is the momentum deposition by waves in the stratosphere,  $A$  is the mean eddy kinetic energy in the troposphere,  $T_{eq}$  is the mean surface temperature at the equator,  $T_p$  is the mean surface temperature of the polar regions, and  $J$  is the mean jet stream velocity.

## 2. Model Description

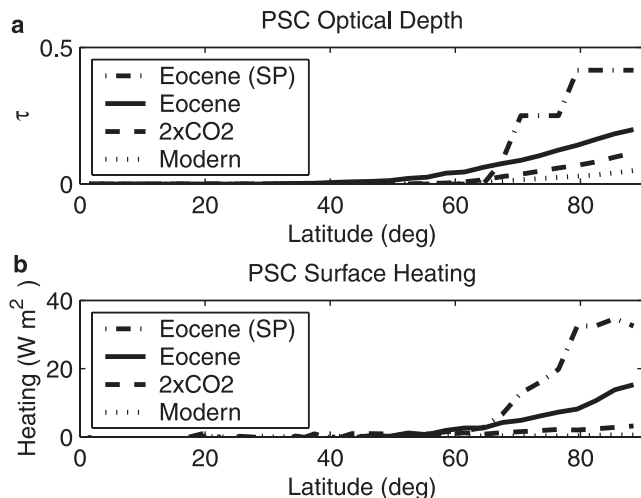
[9] We use a two-dimension climate model with high vertical resolution (15 mb) to estimate the potential strength of this feedback mechanism, to explore its timing over the seasonal cycle, and to consider why this mechanism might be missed in GCM studies. The model is a modified version of the two-dimensional (latitude and atmospheric pressure) energy balance model presented in *Kirk-Davidoff et al.* [1999], and uses the radiative transfer codes of *Chou and Suarez* [1999] for solar radiation and *Chou et al.* [2001] for terrestrial radiation. In the runs described here, surface static stability is set by requiring the isentrope leaving the equator to intersect the polar tropopause. Thus, polar static stability tends to decrease with decreasing EPTD. This is important, because additional model runs show that the PSC feedback is inversely proportional to polar static stability. Tropospheric temperatures are set to satisfy an assumed distribution of potential vorticity in the mid-latitude troposphere, and to lie along moist adiabats in the tropics. Radiative contributions from climatological clouds, relative humidity, and ozone are included. Stratospheric water vapor mixing ratio is set equal to the saturation water vapor mixing ratio for the temperature and pressure of the coldest level of the atmosphere above the equator, thus allowing determination of  $\partial q / \partial T_t$ . Two steps in the feedback loop described above are imposed in the model. We make no claim that the manner in which this is done is true to the real atmosphere. Rather, we are using the model to make an estimate of the strength of that feedback loop assuming that it functions in the simplest conceivable way. First, stratospheric overturning is made to depend linearly on the meridional temperature gradient: the overturning circulation is proportional to the weighted average of the northern and southern equator-to-pole temperature gradient, with the northern hemisphere weighted twice as heavily as the southern hemisphere, to reflect the observed dependence, with the constant of proportionality set such that the overturning yields a heating rate of  $-0.5 \text{ K d}^{-1}$  at the equator and  $+0.5 \text{ K d}^{-1}$  at the poles when  $T_{eq} - T_p = 45 \text{ K}$ , as in the observed modern atmosphere [*Eluszkiewicz et al.*, 1996]. Second, PSC optical depth at each grid point in the stratosphere is made to depend on the excess of stratospheric water vapor mixing ratio (wvmr) above saturation wvmr, by assuming a fixed number density of spherical cloud particles. Particle radius is then given by  $R = \left( \frac{3 r^*}{4 \pi n \rho_i} \right)^{1/3}$ , where  $R$  is the radius,  $r^*$  is the wvmr,  $n$  is the particle number density (assumed equal to  $1 \text{ cm}^{-3}$ ), and  $\rho_i$  is the density of ice. Given particle radius and cloud water mixing ratio, the radiative transfer code computes the cloud optical depth.

## 3. Results and Discussion

[10] Although the model can be used to derive the surface temperature, we will concentrate on results from runs in which the surface temperature is prescribed. By prescribing the surface temperature, we can isolate the response of the PSCs and their heating to changes in the surface temperature distribution. In the fixed surface temperature runs, surface temperature is set as follows:

$$T = T_{eq} - T_p(1.0 - \cos \phi) - \Delta T_s \cos(\pi \frac{m-2}{6}) \sin \phi | \sin \phi | \quad (3)$$

where  $T_{eq}$  is the temperature at the equator,  $T_p$  is the temperature difference between the equator and the poles,  $\Delta T_s$  is one-half the



**Figure 1.** PSC optical depth (a) and surface heating (b) for Modern,  $2 \times \text{CO}_2$ , and Eocene surface temperature distributions. Eocene (SP) denotes Eocene surface temperature distributions with PSCs inserted as in *Sloan and Pollard* [1998].

amplitude of the seasonal cycle at the poles,  $m$  is the month of the year, and  $\phi$  is latitude.

[11] In the model, stratospheric water vapor increases as the EPTD decreases, leading to an increase in PSC optical depth and surface heating (See Table 1). The PSC effect is tested by running the model twice for each temperature distribution, once with PSC physics switched on, and once off. Surface heating and outgoing radiation are then differenced to obtain the change due to PSCs. In addition, for the Eocene temperature distribution, an additional model run was made, denoted “Eocene (SP)”, in which PSCs are imposed as in *Sloan and Pollard* [1998], occurring at 60 mb with unity optical depth in all regions of polar night. Results from the no PSC case were subtracted from this run to yield the surface warming and change in outgoing radiation due to the PSCs imposed in this way. Surface heating ( $F_p$ ) includes both the direct thermal radiative heating due to the PSCs, and the reduced convective heat losses the model predicts (due to radiative heating and stabilization of the model troposphere). The latter, convective effect, is far larger (90% of the whole).

[12] For the “Eocene” case, with polar surface temperatures averaging 285 K, the annual average PSC induced heating is substantial ( $15 \text{ W m}^{-2}$  at the poles), though smaller than “Eocene (SP)” case ( $36 \text{ W m}^{-2}$ ) (Figure 1). As in *Sloan and Pollard* [1998], heating due to PSCs extends equatorward from the PSCs themselves, because radiative cooling of the free troposphere is reduced below the PSCs, and this reduction implies a reduced loss of heat from the surface below and equatorward of the reduction. In the “ $2 \times \text{CO}_2$ ” run, meant to simulate a greenhouse induced warming of the modern climate, our model predicts a modest increase in PSC optical depth, and a very small net heating, since the PSC warming effect is reduced in a more stable polar atmosphere, where the model predicts little convective loss of heat from the surface to the free troposphere. Changes in outgoing longwave radiation (OLR, shown in Table 1) confirm that the PSCs act to reduce the meridional heat flux needed to balance an increase in surface temperature. By integrating the change in top-of-atmosphere radiative fluxes, we find that for the “Eocene” case, the effect of including PSCs is to reduce the meridional energy flux required for equilibrium by 0.1 PW, a 10% reduction in the flux through the 60° latitude parallel in each hemisphere.

[13] Because of the simplicity of the model’s treatment of static stability, we do not have high confidence in its prediction of surface temperature changes in response to stratospheric warming.

With that caveat, we note that when surface temperature are allowed to vary, our model predicts a 7 K warming of polar regions due to the inclusion of PSC effects, and a 10 K warming if PSCs are imposed as in *Sloan and Pollard* [1998]. The latter warming is somewhat smaller than the warming found by *Sloan and Pollard* [1998], a difference which may be attributable to the simplicity of our two dimensional model’s surface static stability parameterization, since the PSC warming effect diminishes sharply as surface static stability increases.

[14] Seasonal variations are similar to those imposed by *Sloan and Pollard* [1998]: PSCs are absent when there is non-zero solar heating of the stratosphere (since temperatures are then above the local frost point), and they form in the polar night, when temperatures are able to fall below the frost point. However, our model is also able to simulate short time-scale variations in PSC optical depth. PSC formation causes radiative heating of air below the PSC and cooling above it. This leads to an upward propagation of the cloud from one day to the next. The rapid temperature fluctuations induced by PSC formation suggest that sedimentation of PSC particles must compete with radiative heating effects if they are to remove stratospheric water before the particles evaporate. This might slow the sedimentation of ice particles that in the modern Antarctic stratosphere substantially reduces the water vapor mixing ratio over the course of the winter [*Kelly et al.*, 1989]. In the real atmosphere, these cloud-related heatings and cooling would tend to force circulations which would further feed back on temperature and cloud structure: analysis of such effects will require more detailed modeling.

[15] Averaging over these fluctuations, the PSCs have a strong overall warming effect on the polar stratosphere, preventing temperatures from falling much below the frost point, and thus reducing the meridional gradient of stratospheric temperature.

[16] These results lead us to inquire why GCM studies of equable climates have not predicted PSC formation. Experiments using our model with varying vertical resolution show that the sensitivity of the tropical tropopause temperature to the strength of the stratospheric overturning circulation depends critically on model resolution. For 50 mb resolution (typical for GCMs), a doubling of the tropical stratospheric upwelling from  $0.3 \text{ K d}^{-1}$  to  $0.6 \text{ K d}^{-1}$  results in a cooling of the tropopause temperature by 2.8 K, while for 10 mb resolution, the same doubling leads to a 6.8 K cooling. Thus, coarse vertical resolution in the neighborhood of the tropical tropopause makes model predictions of stratospheric water vapor unreliable. To avoid this problem, while gaining the ability to predict the planetary wave propagation from first principles, we are currently running simulations of stratospheric wave propagation for varying tropospheric climates in a simplified GCM which be run with high vertical resolution in reasonable integration times.

## 4. Conclusions

[17] An inspection of equations 1 and 2, the differential equation for heating due to PSCs at the polar surface, clearly identifies the key quantities that must be determined by observation and interpretation within the atmosphere. Some of these derivatives can now be calculated with the required accuracy. For example,  $\partial F_p / \partial \tau$ ,  $\partial \tau / \partial q$ , and  $\partial \tau / \partial T_{st}$ , are all obtained from radiative transfer theory, laboratory and field data and from thermodynamic considerations. Those terms that define the quantitative response of atmospheric transport to changes in climate forcing such as  $\partial q / \partial T$ , and  $\partial \omega / \partial \text{CO}_2$ , clearly summarize the agenda for future research.

[18] If PSCs are sensitive to increases in  $\text{CO}_2$  concentration, and this feedback is not being simulated by GCMs, then the GCMs used to predict the effects of anthropogenic  $\text{CO}_2$  emissions underestimate the sensitivity of polar climates to increasing  $\text{CO}_2$ . If the overall response of the climate system to projected increases in

CO<sub>2</sub> is at the low end of GCM predictions, or the response of the stratospheric overturning circulation to changes in the EPTD is superlinear, the feedback discussed here would be relevant for the Eocene, but of little consequence for future climate. On the other hand, if the climate system responds strongly to CO<sub>2</sub> increases, and the stratospheric overturning responds directly to the EPTD, current climate models may seriously underestimate the degree of polar warming which will accompany increasing greenhouse gas concentration over the next century as levels approach those of the Eocene.

[19] **Acknowledgments.** The authors thank Dr. Ming-Dah Chou for the use of his radiation codes, and Dr. Raymond Pierrehumbert for his very helpful review. The contributions of DBKD and JGA were supported by NASA grant NAG-1-01095, and NOAA grant 50-SPNA-00042. DPS's work was supported by NSF grant OCE-9733688, and by NOAA grant NA166P1022.

## References

- Bush, A. B. G., and S. G. H. Philander, The late Cretaceous: Simulation with a coupled atmosphere-ocean general circulation model, *Paleoceanography*, *12*, 495–516, 1997.
- Charney, J. G., and P. G. Drazin, Propagation of planetary-scale disturbances from lower into upper atmosphere, *J. Geophys. Res.*, *66*, 83–110, 1961.
- Chou, M.-D., and M. J. Suarez, A solar radiation parameterization (CLIR-AD-SW) for atmospheric studies. NASA Technical Memorandum 104606, vol. 15, 48 pp., 1999. Available at ([http://climate.gsfc.nasa.gov/chou/clirad\\_sw](http://climate.gsfc.nasa.gov/chou/clirad_sw)).
- Chou, M.-D., M. J. Suarez, X.-Z. Liang, and M. M.-H. Yan, A thermal infrared radiation parameterization for atmospheric studies. NASA Technical Memorandum 104606, vol. 19, 65 pp., 2001. Available at ([http://climate.gsfc.nasa.gov/chou/clirad\\_lw](http://climate.gsfc.nasa.gov/chou/clirad_lw)).
- Douglass, R. G., and F. Woodruff, Deep sea benthic foraminifera, in *The Sea*, vol. 1, pp. 1233–1327, edited by C. Emiliani, Wiley-Interscience, New York, 1981.
- Eluszkiewicz, J., D. Crisp, R. Zurek, L. Elson, E. Fishbein, L. Froidevaux, J. Waters, R. G. Grainger, A. Lambert, R. Harwood, and G. Peckham, Residual circulation in the stratosphere and lower mesosphere as diagnosed from Microwave Limb Sounder data, *J. Atmos. Sci.*, *53*, 217–240, 1996.
- Farrell, B. F., Equable climate dynamics, *J. Atmos. Sci.*, *47*, 2986–2995, 1990.
- Forster, P. M. D., and K. P. Shine, Stratospheric water vapour changes as a possible contributor to observed stratospheric cooling, *Geophys. Res. Lett.*, *26*, 3309–3312, 1999.
- Fusco, A. C., and M. L. Salby, Interannual variations of total ozone and their relationship to variations of planetary wave activity, *J. Climate*, *12*, 1619–1629, 1999.
- Haynes, P. H., C. J. Marks, M. E. McIntyre, T. G. Shepherd, and K. P. Shine, On the downward control of extratropical diabatic circulations by eddy-induced mean zonal forces, *J. Atmos. Sci.*, *48*, 651–679, 1991.
- Kelly, K. K., A. F. Tuck, D. M. Murphy, M. H. Proffitt, D. W. Fahey, R. L. Jones, D. S. McKenna, M. Loewenstein, J. R. Podolske, S. E. Strahan, G. V. Ferry, K. R. Chan, J. F. Vedder, G. L. Gregory, W. D. Hypes, M. P. McCormick, E. V. Browell, and L. E. Heidt, Dehydration in the lower antarctic stratosphere during late winter and early spring, *J. Geophys. Res.*, *94*, 11,317–11,357, 1989.
- Kirk-Davidoff, D. B., E. J. Hints, J. G. Anderson, and D. W. Keith, The effect of climate change on ozone depletion through changes in stratospheric water vapour, *Nature*, *402*, 401–499, 1999.
- Kirk-Davidoff, D. B., and R. S. Lindzen, An energy balance model based on potential vorticity homogenization, *J. Climate*, *13*, 431–448, 2000.
- Lelieveld, J., P. J. Crutzen, and F. J. Dentener, Changing concentration, lifetime and climate forcing of atmospheric methane, *Tellus*, *50*, 128–150, 1998.
- Lindzen, R. S., and B. F. Farrell, The role of polar regions in global climate, and a new parameterization of global heat transport, *Mon. Wea. Rev.*, *108*, 2064–2079, 1980.
- Lindzen, R. S., and W. W. Pan, A note on orbital control of equator-pole heat fluxes, *Clim. Dyn.*, *10*, 49–57, 1994.
- Oltmans, S. J., and D. J. Hofmann, Increase in lower-stratospheric water-vapor at a mid-latitude northern-hemisphere site from 1981 to 1994, *Nature*, *374*, 146–149, 1995.
- Pearson, P. N., P. W. Ditchfield, J. Singano, K. G. Harcourt-Brown, C. J. Nicholas, R. K. Olsson, N. J. Shackleton, and M. A. Hall, Warm tropical sea surface temperatures in the Late Cretaceous and Eocene epochs, *Nature*, *413*, 481–487, 2001.
- Rind, D., Latitudinal temperature gradients and climate change, *J. Geophys. Res.*, *103*, 5943–5971, 1998.
- Rind, D., M. Chandler, P. Lonergan, and J. Lerner, Climate change and the middle atmosphere 5. Paleostratosphere in cold and warm climates, *J. Geophys. Res.*, *106*, 20,195–20,212, 2001.
- Schrag, D. P., Effects of diagenesis on the isotopic record of late paleogene tropical sea surface temperatures, *Chem. Geol.*, *161*, 215–224, 1999.
- Shindell, D. T., D. Rind, and P. Lonergan, Increased polar stratospheric ozone losses and delayed eventual recovery owing to increasing greenhouse-gas concentrations, *Nature*, *392*, 589–592, 1998.
- Sherwood, S. C., A microphysical connection between biomass burning, cumulus clouds, and stratospheric moisture, submitted to *J. Geophys. Res.*, 2001.
- Sloan, L. C., and E. J. Barron, A comparison of Eocene climate model results to quantified paleoclimatic interpretations, *Paleogeogr. Paleoclimatol. Paleocool.*, *93*, 183–202, 1992.
- Sloan, L. C., and D. Pollard, Polar stratospheric clouds: A high latitude warming mechanism in an ancient greenhouse world, *Geophys. Res. Lett.*, *25*, 3517–3520, 1998.
- Toggweiler, J. R., and H. Bjornsson, Drake Passage and paleoclimate, *J. Quat. Sci.*, *15*, 319–328, 2000.
- Weinstock, E. M., E. J. Hints, D. B. Kirk-Davidoff, J. G. Anderson, A. E. Andrews, R. L. Herman, C. R. Webster, M. Loewenstein, J. R. Podolske, and T. P. Bui, Constraints on the seasonal cycle of stratospheric water vapor using in situ measurements from the ER-2 and a CO photochemical clock, *J. Geophys. Res.*, in press, 2001.
- Zachos, J. C., L. D. Stott, and K. C. Lohmann, Evolution of early Cenozoic marine temperatures, *Paleoceanography*, *9*, 353–387, 1994.

D. B. Kirk-Davidoff, D. P. Schrag, and J. G. Anderson, Department of Earth and Planetary Sciences, Harvard University, 20 Oxford St., Cambridge, MA 02144, USA. ([davidoff@huarp.harvard.edu](mailto:davidoff@huarp.harvard.edu); [schrag@eps.harvard.edu](mailto:schrag@eps.harvard.edu); [anderson@huarp.harvard.edu](mailto:anderson@huarp.harvard.edu))