to 1 with respect to M_s. Similarly, J. B. Rundle [*J. Geophys. Res.* **94**, 12377 (1989)], assuming the validity of Eq. 4b, used Eqs. 1a, 1b, 2b, and 3 to derive $b = 1.5$ for large earthquakes, which has been recently confirmed observationally (*5*). Other studies that have developed Eqs. 1 through 4 to serve one purpose or another include T. C. Hanks, *ibid.* **84**, 2235 (1979); D. J. Andrews, *ibid.* **85**, 3867 (1980); K. Aki, in *Earthquake Prediction: An International Review*, D. W. Simpson and P. G. Richards, Eds. (American Geophysical Union, Washington, DC, 1981), pp. 566–574; G. King, *Pure Appl. Geophys.* **121**, 761 (1983); D. Turcotte, *ibid.* **131**, 171 (1989); A. Frankel, *J. Geophys. Res.* **96**, 6291 (1991). Readers who execute the algebra to reach Eq. 4b will note that $N, \Delta N \sim (\ell w)^{-1} w^{-1}$. Because w is customarily taken as a constant, this large-earthquake scaling reduces in fact to reciprocal fault-length scaling.

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Effects of Aerosol from Biomass Burning on the Global Radiation Budget

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An analysis is made of the likely contribution of smoke particles from biomass burning to the global radiation balance. These particles act to reflect solar radiation directly; they also can act as cloud condensation nuclei, increasing the reflectivity of clouds. Together these effects, although uncertain, may add up globally to a cooling effect as large as 2 watts per square meter, comparable to the estimated contribution of sulfate aerosols. Anthropogenic increases of smoke emission thus may have helped weaken the net greenhouse warming from anthropogenic trace gases.

Atmospheric aerosol particles have a noticeable effect on solar radiation and are sufficiently widespread for this effect to have implications for climate. However, their relatively short and irregular lifetimes have made it difficult to quantify their contribution to regional and global radiation budgets. Tropospheric aerosols (which have lifetimes of less than a week and an extremely patchy horizontal spatial distribution) may, in total, cool climate through their direct radiative effect by an amount comparable to the warming expected from a doubling of CO_2 (*1*).

Sulfate has been recognized as the dominant contributor to tropospheric aerosols over and near industrial continental areas (*2*). Smoke aerosol may have a similar role over and near tropical continental areas. The production of both sulfate and smoke aerosol from human activities is expected to have greatly intensified during the last century.

Recent estimates of the contribution of anthropogenic emissions of sulfate aerosol to climate change have been based on the combined approaches of biogeochemical cycling, atmospheric radiative transfer theory, and climate modeling (*2*). The direct and indirect contributions of the anthropogenic sulfate aerosol to the atmospheric radiation balance appear to be comparable. They are similar in magnitude and opposite in sign to the effects of the increase in atmospheric greenhouse gases over the last century. The global effect of aerosols from smoke adds further to this effect. Thus, the increase of aerosols over the last century may explain why the warming observed over this period is at the lower limit of that indicated by climate models (*3*) from the increased levels of greenhouse gases.

The global mass of smoke, M, in the atmosphere depends on the total rate, R, of biomass burning, the fraction, f, of burned material that goes into smoke, and the lifetime, t_s, of the smoke in the atmosphere

$$M = fRt_s \qquad (1)$$

Most (more than 70%) of the burned biomass is in the tropics, with contributions from burning of forests for shifting agriculture and burning of virgin forest cleared for colonization, savannas, fuel wood, and agricultural wastes (*4*). Recent estimates of tropical biomass burning from the Food and Agriculture Organization statistics for the period 1975 to 1980 (*5*) suggest that burning in savannas dominates that in forests, in contrast to earlier estimates (*4*). The annual carbon burning from deforestation, savanna burning, and shifting agriculture is estimated to be 1300 to 3300 Tg of C (*5*) (1 Tg = 10^{12} g). An estimate of the contributions of fuel wood, agricultural wastes, and charcoal is 1000 to 2000 Tg of C per year (*6*), or, in sum, the C burned per year globally is likely 2600 to 5000 Tg.

Measurements of the fraction of smoke produced (*7–9*) suggest that

$$f = 0.03 \text{ g of smoke per gram of C} \qquad (2)$$

Thus, for $R = 3800$ Tg of C per year, 114 Tg of smoke is produced per year. This value is comparable to the global annual anthropogenic sulfate production of 110 Tg per year (*2*). The smoke has approximately the same distribution of particle sizes as sulfate aerosols. In particular, the bulk of the mass and effectively all of the optical effects are associated with or determined by particles of this size in the aerosol accumulation mode with a mass-averaged radius of around 0.3 μm. Because these particles include hygroscopic components, as do sulfate particles, they should have a comparable lifetime. For a lifetime of 6 days (*2*), Eq. 1 implies that the global smoke loading is $M = 3.7 \times 10^{-3}$ g m^{-2}. The optical thickness of an aerosol layer is

$$\tau = \tau_a + \tau_s \qquad (3)$$

where τ_a is the absorption optical depth and τ_s is the scattering optical depth. Optical

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properties of smoke have been estimated for a number of types of forest fires from remote sensing (*10*) and by direct sampling (*8*). Single scattering albedos, ω, characterize the ratio of scattering to absorption plus scattering for a thin aerosol layer. The data support a value for ω of 0.85 to 0.90 for dry smoke aerosols and 0.90 to 0.95 for smoke aerosols that have adjusted to ambient humidity conditions by collecting water molecules.

Total mass extinction cross sections of dry smoke have been computed as about 4.5 $m^2\ g^{-1}$ (*10*) and measured as 5 $m^2\ g^{-1}$ on average (*8*). Charlson *et al.* (*2*) measured extinctions for sulfate aerosol of 5 $m^2\ g^{-1}$, but correction for ambient humidity yielded 8.5 $m^2\ g^{-1}$. With a similar humidity correction, we infer a value of 8 $m^2\ g^{-1}$ for τ_s for smoke. With Eq. 3, this value implies that the global average optical depth due to smoke aerosol is about 0.03, or about the same as the sulfate aerosol estimate of Charlson *et al.* (*2*). This value may be quite a bit less than a total tropospheric aerosol attenuation (*11*), but smoke can be dominant over tropical continents during times of burning. An optical depth of 0.01 has been estimated (*12*) for a scattering coefficient half as great and a loading two-thirds as great as we inferred. The treatment of Charlson *et al.* (*2*) can be used to assess the climatic role of biomass smoke. A simple radiation model of an optically thin aerosol underlying the atmosphere and over the surface gives a change of planetary albedo α_p of

$$\Delta\alpha_p = [2T^2a(1 - A_c)(1 - \alpha_s)^2\beta]\tau_s \qquad (4)$$

when T_a is the atmosphere clear-sky transmittance, A_c is the fraction of clouds, α_s is the surface albedo, and β is the fraction of radiation scattered upward by the aerosol layer of optical depth τ_s. The term in brackets has a numerical value of about 0.13 over ocean and 0.09 over land (*13*); thus, for smoke aerosol loading given by Eq. 3, $\Delta\alpha_p = 0.003$, in which case the smoke leads to an additional reflection of 1.0 W m^{-2}.

We have neglected absorption of solar radiation by smoke aerosols. Such absorption may be important for heating atmospheric layers (*14*), even if net global albedos are little affected. Absorption by aerosols heats the atmosphere relative to the surface and may inhibit moist convection. For an estimated absorption coefficient of 0.7 $m^2\ g^{-1}$ (*8*) and a scattering coefficient of 8 $m^2\ g^{-1}$, the single-scatter albedo is 0.92. For our estimate of smoke loading, this absorption coefficient leads to τ_a of about 0.0026 (solar absorption of about 0.5 W m^{-2} allowing for clouds). Correspondingly, because of atmospheric absorption, absorption at the surface is reduced by about 0.3 W m^{-2} or the net heating of the system from smoke absorption is about 0.2 W m^{-2}.

Aerosols acting as cloud condensation nuclei (CCN) change the radiative properties of clouds (*15–17*). The number concentration of cloud droplets in stable cloud layers is proportional to the concentration N of aerosol CCN, and the factor of proportionality is typically in the range 0.7 to 0.8 (*7, 16*). The derivative of cloud albedo, α_c, with respect to the logarithm of cloud droplet concentration, is typically near its maximum value of about 0.083 over typical cloud thicknesses (*2*), so we estimate that

$$\Delta\alpha_c = 0.06\ \Delta N/N \qquad (5)$$

The global average concentration of CCN from smoke is estimated to be 10^2 to 10^3 (*18*). Concentrations over the ocean will be lower. Added to a background of 50 to 100 cm^{-3}, 10 cm^{-3} smoke particles could increase the albedo of marine stratiform clouds by 0.01, corresponding to a cooling of global climate by an additional 1 W m^{-2} (*19*).

Our analysis suggests that the direct radiative effects of smoke aerosol from biomass burning and the indirect effects through clouds may be of equal significance for climate. The global reflection of solar radiation may have been increased by those effects by up to ~2 W m^{-2}, comparable to that from sulfate aerosol. Increased emissions of greenhouse gases over the last century now warm the planet by ~2 W m^{-2} (*3*). Some aspects of the estimated cooling from aerosols may now be exaggerated because otherwise global temperature decreases should have been observed during the last century. The estimated direct effects of smoke aerosol are probably correct within a factor of 2, but the indirect effects could be uncertain by at least twice as much. It seems likely that the net cooling from emissions of anthropogenic aerosols (sulfate plus smoke) over the last century has been at least half that of the anthropogenic greenhouse warming. A near balance between anthropogenic greenhouse warming and aerosol cooling would have implications for the analysis of present temperature trends, particularly in terms of climate sensitivity and strategies for reducing future greenhouse warming (*20*).

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Oxygen Isotopes in Biogenic Silica: Global Changes in Ocean Temperature and Isotopic Composition

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A record of oxygen isotopes in biogenic silica from a deep-sea sediment core from the Southern Ocean reveals that marine diatoms retain their primary isotopic composition after burial. As a result, the marine diatom record can be combined with data on coexisting planktonic foraminifera to monitor past surface temperature and isotopic composition of seawater. The coupling of these two records allows the solution of two paleotemperature equations for each core interval. Data from a South Atlantic core show that the average $\delta^{18}O$ during the glacial period at this site was higher by about 1.3 per mil than average Holocene values, and that average glacial-age temperatures were not significantly different from average Holocene values.

The oxygen isotopic composition of foraminiferal calcite ($\delta^{18}O_c$) depends on the temperature and the isotopic composition of seawater ($\delta^{18}O_w$) from which it was deposited (*1*). Differentiation between these two components of the Pleistocene marine $\delta^{18}O$ record has been a long-standing problem because shifts in both temperature and mean oceanic $\delta^{18}O_w$ (a function of the amount of isotopically light water stored on continents in the form of ice caps) are crucial boundary conditions for climate models. One approach has been to estimate global ice volume by analysis of benthic foraminifera recovered from deep-sea sediments, with the assumption that the temperature of the bottom waters in which these foraminifera grew remained relatively constant (*2–4*). However, on the basis of ice sheet dynamics (*5*) and detailed comparisons between the deep-sea record and the elevation of coral terraces, it has become clear that the benthic $\delta^{18}O_c$ record does not trace ice volume accurately (*6–8*). Another, less inferential approach is to measure the isotopic composition of a coexisting phase in which oxygen isotopes are fractionated with a different temperature dependency. In this report we show that marine biogenic silica may be a suitable phase for such a comparison and describe a record of sea-surface temperature (SST) and $\delta^{18}O_w$ from the South Atlantic.

Marine biogenic silica is composed primarily of diatoms (marine algae) that deposit internal silica frustules. Silicification occurs only in the uppermost layer of the ocean, because of the light requirements for photosynthesis. Therefore, unlike planktonic foraminifera, which have the capacity to migrate through the water column, diatoms represent an ideal recorder of SST and the isotopic composition of seawater, provided that they deposit their silica with a known isotopic fractionation. Early attempts to measure the isotopic composition of biogenic silica ($\delta^{18}O_{Si}$) for paleoceanographic purposes (*9–12*) were hampered by difficulties in obtaining a clean diatom fraction, poor reproducibility due to isotopic exchange of oxygen between silica and hydration water during laboratory preparations, and initial results which suggested that the slope of the $\delta^{18}O_{Si}$-temperature relation was similar to that for carbonate.

More recent studies have suggested that these problems have been overcome. Diatoms in sediment may now be cleaned efficiently by differential settling and sieving combined with acid cleaning (*13*). Juillet-Leclerc and Labeyrie reported (*14*) an isotopic exchange technique that allows control and accurate calculation of the isotopically unstable (exchangeable) fraction of oxygen atoms in the silica structure. Measurement of diatoms from surface sediments with the use of this technique led to calibration of a paleotemperature equation (*14*):

$$t = 17.2 - 2.4(\delta^{18}O_{Si} - \delta^{18}O_w - 40) - 0.2(\delta^{18}O_{Si} - \delta^{18}O_w - 40)^2 \tag{1}$$

where t is temperature in degrees celsius. The slope of this equation is significantly different from that of the carbonate paleotemperature equation. This equation was calibrated between 1.5° and 24°C, based on samples from the Gulf of California for the high-temperature range. However, because effects of local upwelling were ignored in the tropical core sites, the equation yields temperatures that are too high for more polar sites. For example, measured $\delta^{18}O_{Si}$ in the top of core RC13-269 (South Atlantic) predicts a silicification temperature of 8.8°C, whereas the observed SST over the core site has never exceeded 4°C. The same applies for the top of the nearby core RC13-271. Therefore, we suggest that for high-latitude cores, it is preferable to use a calibration based only on the low-temperature data points reported in (*14*):

$$t = 11.03 - 2.03(\delta^{18}O_{Si} - \delta^{18}O_w - 40) \tag{2}$$

The slopes of Eqs. 1 and 2 are similar, and the recalibration changes primarily the intercept. The effect of temperature variability on $\delta^{18}O_{Si}$ is roughly twice that for carbonate, although more data points are needed to refine the $\delta^{18}O_{Si}$ paleotemperature equation. The difference in slopes suggests that it is possible to obtain a past record of both $\delta^{18}O_w$ and SST from diatom silica and planktonic foraminifera in a deep-sea core provided that both plankton groups deposited their solid phase at the same depth and time of year.

This condition is apparently satisfied in the Southern Ocean. Data from South Atlantic core tops suggest that the $\delta^{18}O$ of the planktonic foraminifer *Neogloboquadrina pachyderma* is close to equilibrium with surface water (*15*), and sediment trap studies indicate that the primary flux of *N. pachyderma* occurs within the seasonal bloom of diatoms (*16*). Therefore, the present diatomaceous ooze belt south of the Antarctic Polar Front (APF) is an appropriate region for paired time series of diatom and foraminiferal $\delta^{18}O$. We chose to analyze the siliceous core RC13-271 (51°59'S 04°31'E; water depth, 3634 m) for $\delta^{18}O_{Si}$ because it contains sufficient numbers of *N. pachyderma* for isotopic analysis and extends through the last glacial period with high resolution.

The 20- to 60-μm fraction of diatoms from 34 levels of core RC13-271 were cleaned and inspected with a microscope to ensure the absence of contaminant phases. The isotopic exchange procedure for $\delta^{18}O_{Si}$ analyses followed that in (*14*) with one modification: we used water with $\delta^{18}O$ of 39 per mil, close to the solid silica value, in the exchange reaction. This procedure eliminates the uncertainty involved with varying amounts of exchangeable oxygen in the silica structure. Fluorination was performed at the Weizmann Institute, and the

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