VOLCANIC ERUPTIONS AND CLIMATE

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Abstract. Volcanic eruptions are an important natural cause of climate change on many timescales. A new capability to predict the climatic response to a large tropical eruption for the succeeding 2 years will prove valuable to society. In addition, to detect and attribute anthropogenic influences on climate, including effects of greenhouse gases, aerosols, and ozone-depleting chemicals, it is crucial to quantify the natural fluctuations so as to separate them from anthropogenic fluctuations in the climate record. Studying the responses of climate to volcanic eruptions also helps us to better understand important radiative and dynamical processes that respond in the climate system to both natural and anthropogenic forcings. Furthermore, modeling the effects of volcanic eruptions helps us to improve climate models that are needed to study anthropogenic effects. Large volcanic eruptions inject sulfur gases into the stratosphere, which convert to sulfate aerosols with an *e*-folding residence time of about 1 year. Large ash particles fall out much quicker. The radiative and chemical effects of this aerosol cloud produce responses in the climate system. By scattering some solar radiation back to space, the aerosols cool the surface, but by absorbing both solar and terrestrial radiation, the aerosol layer heats the stratosphere. For a tropical eruption this heating is larger in the tropics than in the high latitudes, producing an enhanced pole-to-equator temperature gradient, especially in winter. In the Northern Hemisphere winter this enhanced gradient produces a stronger polar vortex, and this stronger jet stream produces a characteristic stationary wave pattern of tropospheric circulation, resulting in winter warming of Northern Hemisphere continents. This indirect advective effect on temperature is stronger than the radiative cooling effect that dominates at lower latitudes and in the summer. The volcanic aerosols also serve as surfaces for heterogeneous chemical reactions that destroy stratospheric ozone, which lowers ultraviolet absorption and reduces the radiative heating in the lower stratosphere, but the net effect is still heating. Because this chemical effect depends on the presence of anthropogenic chlorine, it has only become important in recent decades. For a few days after an eruption the amplitude of the diurnal cycle of surface air temperature is reduced under the cloud. On a much longer timescale, volcanic effects played a large role in interdecadal climate change of the Little Ice Age. There is no perfect index of past volcanism, but more ice cores from Greenland and Antarctica will improve the record. There is no evidence that volcanic eruptions produce El Niño events, but the climatic effects of El Niño and volcanic eruptions must be separated to understand the climatic response to each.

1. INTRODUCTION

Volcanism has long been implicated as a possible cause of weather and climate variations. Even 2000 years ago, Plutarch and others [*Forsyth*, 1988] pointed out that the eruption of Mount Etna in 44 B.C. dimmed the Sun and suggested that the resulting cooling caused crops to shrivel and produced famine in Rome and Egypt. No other publications on this subject appeared until Benjamin Franklin suggested that the Lakagigar eruption in Iceland in 1783 might have been responsible for the abnormally cold summer of 1783 in Europe and the cold winter of 1783–1784 [*Franklin*, 1784]. *Humphreys* [1913, 1940] associated cooling events after large volcanic eruptions with the radiative effects of the stratospheric *aerosols* but did not have a sufficiently long or horizontally extensive temperature database to quantify the effects. (Terms in italic are defined in the glossary, which follows the main text.) *Mitchell* [1961] was the first to conduct a superposed epoch analysis, averaging the effects of several eruptions to isolate the volcanic effect from other presumably random fluctuations. He only looked at 5-year average periods, however, and did not have a very long temperature record. Several previous reviews of the effects of volcanoes on climate include *Lamb* [1970], *Toon and Pollack* [1980], *Toon* [1982], *Ellsaesser* [1983], *Asaturov et al.* [1986], *Kondratyev* [1988], *Robock* [1989, 1991], and *Kondratyev and Galindo* [1997]. Past theoretical studies of the radiative effects include *Pollack et al.* [1976], *Harshvardhan* [1979], *Hansen et al.* [1992], and *Stenchikov et al.* [1998]. The work of H. H. Lamb, in fact, was extremely influential in the modern study of the impact of volcanic eruptions on climate [*Kelly et al.*, 1998]. Since these reviews, a deeper and more complex understanding of the impacts of volcanic eruptions on weather and climate has resulted, driven by the many

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TABLE 1. Major Volcanic Eruptions of the Past 250 Years

Volcano	*Year of Eruption*	*VEI*	*DVI/E_{max}*	*IVI*
Grimsvotn [Lakagigar], Iceland	1783	4	2300	0.19
Tambora, Sumbawa, Indonesia	1815	7	3000	0.50
Cosiguina, Nicaragua	1835	5	4000	0.11
Askja, Iceland	1875	5	1000	0.01*
Krakatau, Indonesia	1883	6	1000	0.12
Okataina [Tarawera], North Island, New Zealand	1886	5	800	0.04
Santa Maria, Guatemala	1902	6	600	0.05
Ksudach, Kamchatka, Russia	1907	5	500	0.02
Novarupta [Katmai], Alaska, United States	1912	6	500	0.15
Agung, Bali, Indonesia	1963	4	800	0.06
Mount St. Helens, Washington, United States	1980	5	500	0.00
El Chichón, Chiapas, Mexico	1982	5	800	0.06
Mount Pinatubo, Luzon, Philippines	1991	6	1000	···

The official names of the volcanoes and the volcanic explosivity index (VEI) [*Newhall and Self*, 1982] are from *Simkin and Siebert* [1994]. The dust veil index (DVI/E_{max}) comes from *Lamb* [1970, 1977, 1983], updated by *Robock and Free* [1995]. The ice core volcanic index (IVI) is the average of Northern and Southern Hemisphere values and is represented as optical depth at $\lambda = 0.55$ μm [from *Robock and Free*, 1995, 1996].
*Southern Hemisphere signal only; probably not Askja.

studies of the impact of the 1991 Pinatubo eruption and continuing analyses of the 1982 El Chichón eruption in Mexico.

This paper reviews these new results, including the indirect effect on atmospheric circulation that produces winter warming of the Northern Hemisphere (NH) continents and the new impacts on ozone due to the stratospheric presence of anthropogenic chlorine. A better understanding of the impacts of volcanic eruptions has important applications in a number of areas. Attribution of the warming of the past century to anthropogenic greenhouse gases requires assessment of other causes of climate change during the past several hundred years, including volcanic eruptions and solar variations. After the next major eruption, new knowledge of the indirect effects on atmospheric circulation will allow better seasonal forecasts, especially for the NH in the winter. The impacts of volcanic eruptions serve as analogs, although imperfect ones, for the effects of other massive aerosol loadings of the atmosphere, including meteorite or comet impacts or nuclear winter.

The largest eruptions of the past 250 years (Table 1) have each drawn attention to the atmospheric and potential climatic effects because of their large effects in the English-speaking world. (*Simkin et al.* [1981] and *Simkin and Siebert* [1994] provide a comprehensive list of all known volcanoes and their eruptions.) The 1783 eruption in Iceland produced large effects all that summer in Europe [*Franklin*, 1784; *Grattan et al.*, 1998]. The 1815 Tambora eruption produced the "year without a summer" in 1816 [*Stommel and Stommel*, 1983; *Stothers*, 1984; *Robock*, 1984a, 1994; *Harington*, 1992] and inspired the book *Frankenstein* [*Shelley*, 1818]. The most extensive study of the impacts of a single volcanic eruption was carried out by the Royal Society, examining the 1883 Krakatau eruption, in a beautifully produced volume including watercolors of the volcanic sunsets near London [*Symons*, 1888; *Simkin and Fiske*, 1983]. This was probably the loudest explosion of historic times, and the book includes color figures of the resulting pressure wave's four circuits of the globe as measured by microbarographs. The 1963 Agung eruption produced the largest stratospheric dust veil in more than 50 years and inspired many modern scientific studies. While the Mount St. Helens eruption of 1980 was very explosive, it did not inject much sulfur into the stratosphere. Therefore it had very small global effects [*Robock*, 1981a]. Its tropospheric effects lasted only a few days [*Robock and Mass*, 1982; *Mass and Robock*, 1982], but it occurred in the United States and so received much attention. Quantification of the size of these eruptions is difficult, as different measures reveal different information. For example, one could examine the total mass ejected, the explosiveness, or the sulfur input to the stratosphere. The limitations of data for each of these potential measures, and a description of indices that have been produced, are discussed later.

Volcanic eruptions can inject into the stratosphere tens of *teragrams* of chemically and microphysically active gases and solid aerosol particles, which affect the Earth's radiative balance and climate, and disturb the stratospheric chemical equilibrium. The volcanic cloud forms in several weeks by SO_2 conversion to *sulfate* aerosol and its subsequent microphysical transformations [*Pinto et al.*, 1989; *Zhao et al.*, 1995]. The resulting cloud of sulfate aerosol particles, with an *e-folding decay time* of approximately 1 year [e.g., *Barnes and Hoffman*, 1997], has important impacts on both shortwave and longwave radiation. The resulting disturbance to the Earth's radiation balance affects surface temperatures through direct radiative effects as well as through indirect effects on the atmospheric circulation. In cold regions of the stratosphere these aerosol particles also serve as surfaces for *heterogeneous chemical reactions* that liberate chlorine to destroy ozone in the same way that water and nitric acid aerosols in polar stratospheric clouds produce the seasonal Antarctic ozone hole.

In this paper I first briefly summarize volcanic inputs to the atmosphere and review our new understanding of the radiative forcing of the climate system produced by volcanic aerosols. Next, I briefly review the results of new analyses of ice cores, since they give information about the record of past volcanism, and compare these new records to past analyses. The effects of eruptions on the local diurnal cycle are reviewed. Summer cooling and winter warming from large explosive eruptions are then explained. The impacts of volcanic eruptions on decadal- and century-scale climate changes, and their contributions to the *Little Ice Age* and their relative contribution to the warming of the past century, are next discussed. Then, I show that the simultaneous occurrence of the 1982 El Niño and the El Chichón eruption was just a coincidence and that it was not evidence of a cause and effect relationship. Finally, the impacts of volcanic eruptions on stratospheric ozone are briefly reviewed.

2. VOLCANIC INPUTS TO THE ATMOSPHERE

Volcanic eruptions inject several different types of particles and gases into the atmosphere (Plate 1). In the past, it was only possible to estimate these volatile inputs based on measurements from active, but not explosive, eruptions and remote sensing of the resulting aerosol clouds from lidar, radiometers, and satellites. The serendipitous discovery of the ability of the *Total Ozone Mapping Spectrometer* (*TOMS*) instrument to monitor SO_2 [e.g., *Bluth et al.*, 1992], however, has given us a new tool to directly measure stratospheric injection of gases from eruptions.

The major component of volcanic eruptions is *magmatic material*, which emerges as solid, *lithic* material or solidifies into large particles, which are referred to as ash or tephra. These particles fall out of the atmosphere very rapidly, on timescales of minutes to a few weeks in the troposphere. Small amounts can last for a few months in the stratosphere but have very small climatic impacts. *Symons* [1888], after the 1883 Krakatau eruption, and *Robock and Mass* [1982], after the 1980 Mount St. Helens eruption, showed that this temporary large atmospheric loading reduced the amplitude of the diurnal cycle of surface air temperature in the region of the tropospheric cloud. These effects, however, disappear as soon as the particles settle to the ground. When an eruption column still laden with these hot particles descends down the slopes of a volcano, this *pyroclastic flow* can be deadly to those unlucky enough to be at the base of the volcano. The destruction of Pompeii and Herculaneum after the 79 A.D. Vesuvius eruption is the most famous example.

Volcanic eruptions typically also emit gases, with H_2O, N_2, and CO_2 being the most abundant. Over the lifetime of the Earth these gases have been the main source of the planet's atmosphere and ocean, after the primitive atmosphere was lost to space. The water has condensed into the oceans, the CO_2 has been changed by plants into O_2, with some of the C turned into fossil fuels. Of course, we eat the plants and the animals that eat the plants, we drink the water, and we breathe the oxygen, so each of us is made of volcanic emissions. The atmosphere is now mainly composed of N_2 (78%) and O_2 (21%), both of which had sources in volcanic emissions.

Of these abundant gases, both H_2O and CO_2 are important greenhouse gases, but their atmospheric concentrations are so large (even for CO_2 at only about 370 ppm but growing) that individual eruptions have a negligible effect on their concentrations and do not directly impact the greenhouse effect. Rather, the most important climatic effect of explosive volcanic eruptions is through their emission of sulfur species to the stratosphere, mainly in the form of SO_2 [*Pollack et al.*, 1976; *Newhall and Self*, 1982; *Rampino and Self*, 1984] but possibly sometimes as H_2S [*Luhr et al.*, 1984; *Ahn*, 1997]. These sulfur species react with OH and H_2O to form H_2SO_4 on a timescale of weeks, and the resulting H_2SO_4 aerosols produce the dominant radiative effect from volcanic eruptions. *Bluth et al.* [1992], from satellite measurements, estimated that the 1982 El Chichón eruption injected 7 *Mt* of SO_2 into the atmosphere, and the 1991 Pinatubo eruption injected 20 Mt.

Once injected into the stratosphere, the large aerosol particles and small ones being formed by the sulfur gases are rapidly advected around the globe. Observations after the 1883 Krakatau eruption showed that the aerosol cloud circled the globe in 2 weeks [*Symons*, 1888]. Both the 1982 El Chichón cloud [*Robock and Matson*, 1983] and the 1991 Pinatubo cloud [*Bluth et al.*, 1992] circled the globe in 3 weeks. Although El Chichón (17°N) and Pinatubo (15°N) are separated by only 2° of latitude, their clouds, after only one circuit of the globe, ended up separated by 15° of latitude, with the Pinatubo cloud straddling the equator [*Stowe et al.*, 1992] and the El Chichón cloud extending approximately from the equator to 30°N [*Strong*, 1984]. Subsequent dispersion of a stratospheric volcanic cloud depends heavily on the particular distribution of winds at the time of eruption, although high-latitude eruption clouds are seldom transported beyond the midlatitudes of the same hemisphere. For trying to reconstruct the effects of older eruptions, this factor adds a further complication, as the latitude of the volcano is not sufficient information.

The normal residual stratospheric meridional circulation lifts the aerosols in the tropics, transports them poleward in the midlatitudes, and brings them back into the troposphere at higher latitudes on a timescale of 1–2 years [*Trepte and Hitchman*, 1992; *Trepte et al.*, 1993; *Holton et al.*, 1995].

Quiescent continuous volcanic emissions, including

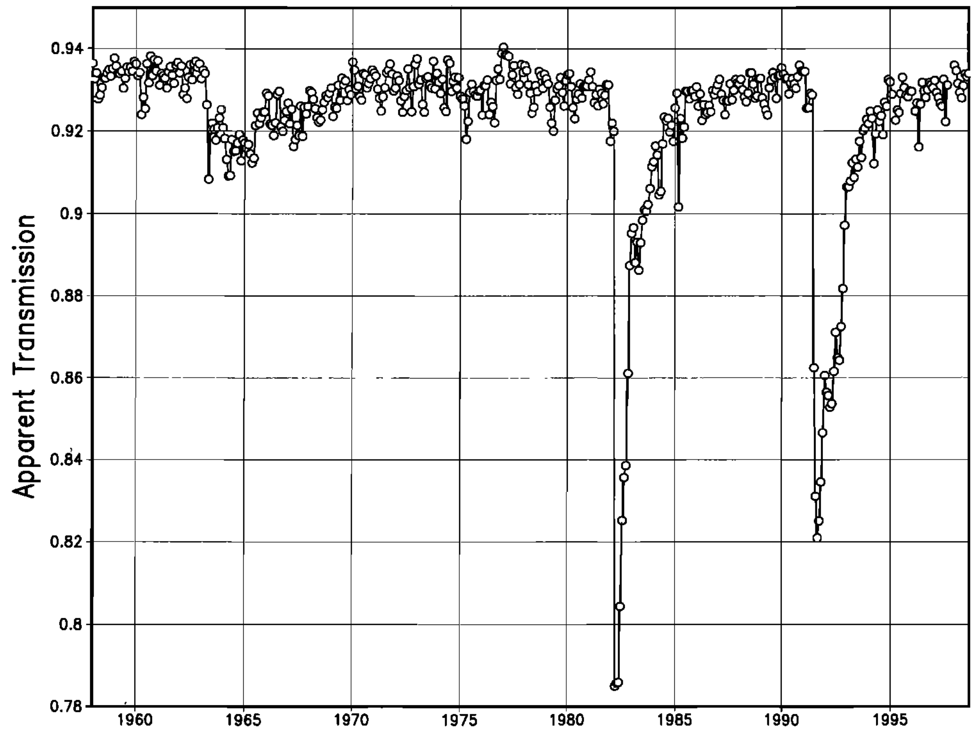


Figure 1. Broadband spectrally integrated atmospheric transmission factor, measured with the *pyrheliometer* shown in Plate 2. *Dutton et al.* [1985] and *Dutton* [1992] describe the details of the calculations, which eliminate instrument calibration and solar constant variation dependence, and show mainly the effects of aerosols. Effects of the 1963 Agung, 1982 El Chichón, and 1991 Pinatubo eruptions can clearly be seen. Years on abscissa indicate January of that year. Data courtesy of E. Dutton.

fumaroles and small episodic eruptions, add sulfates to the troposphere, but their lifetimes there are much shorter than those of stratospheric aerosols. Therefore they are not important for climate change, but they could be if there is a sudden change or a long-term trend in them develops. Global sulfur emission of volcanoes to the troposphere is about 14% of the total natural and anthropogenic emission [*Graf et al.*, 1997] but has a much larger relative contribution to radiative effects. Many volcanic emissions are from the sides of mountains, above the atmospheric boundary layer, and thus they have longer lifetimes than anthropogenic aerosols. Radiative forcing (measured at the surface) from such emissions is estimated to be about -0.2 W m^{-2} for the globe and -0.3 W m^{-2} for the NH, only a little less than anthropogenic effects.

3. RADIATIVE FORCING

Plate 1 indicates the major radiative processes resulting from the stratospheric aerosol cloud from a major volcanic eruption. The most obvious and well-known effect is on solar radiation. Since the sulfate aerosol particles are about the same size as visible light, with a typical effective radius of 0.5 μm, but have a single-scatter albedo of 1, they strongly interact with solar radiation by scattering. Some of the light is backscattered, reflecting sunlight back to space, increasing the net planetary albedo and reducing the amount of solar energy that reaches the Earth's surface. This backscattering is the dominant radiative effect at the surface and results in a net cooling there. Much of the solar radiation is forward scattered, resulting in enhanced downward diffuse radiation that somewhat compensates for a large reduction in the direct solar beam. The longest continuous record of the effects of volcanic eruptions on atmospheric transmission of radiation is the *apparent transmission* record [*Dutton et al.*, 1985; *Dutton*, 1992] from the Mauna Loa Observatory (Plate 2) shown in Figure 1. The effects of the 1963 Agung, 1982 El Chichón, and 1991 Pinatubo eruptions can be clearly seen. Although the Pinatubo eruption produced the largest stratospheric input of the three, the center of the El Chichón cloud went directly over Hawaii, while only the side of the Pinatubo cloud was observed. The Agung cloud was mostly in the Southern Hemisphere, so only the edge was seen in Hawaii. Figure 2 shows separate direct and diffuse radiation measurements, also from Mauna Loa, which show not only the strong reduction of direct radiation by the 1982 El Chichón and 1991 Pinatubo eruptions but also the compensating increase (of slightly smaller amplitude) in the diffuse radiation.

The effect on solar radiation is so strong that it can

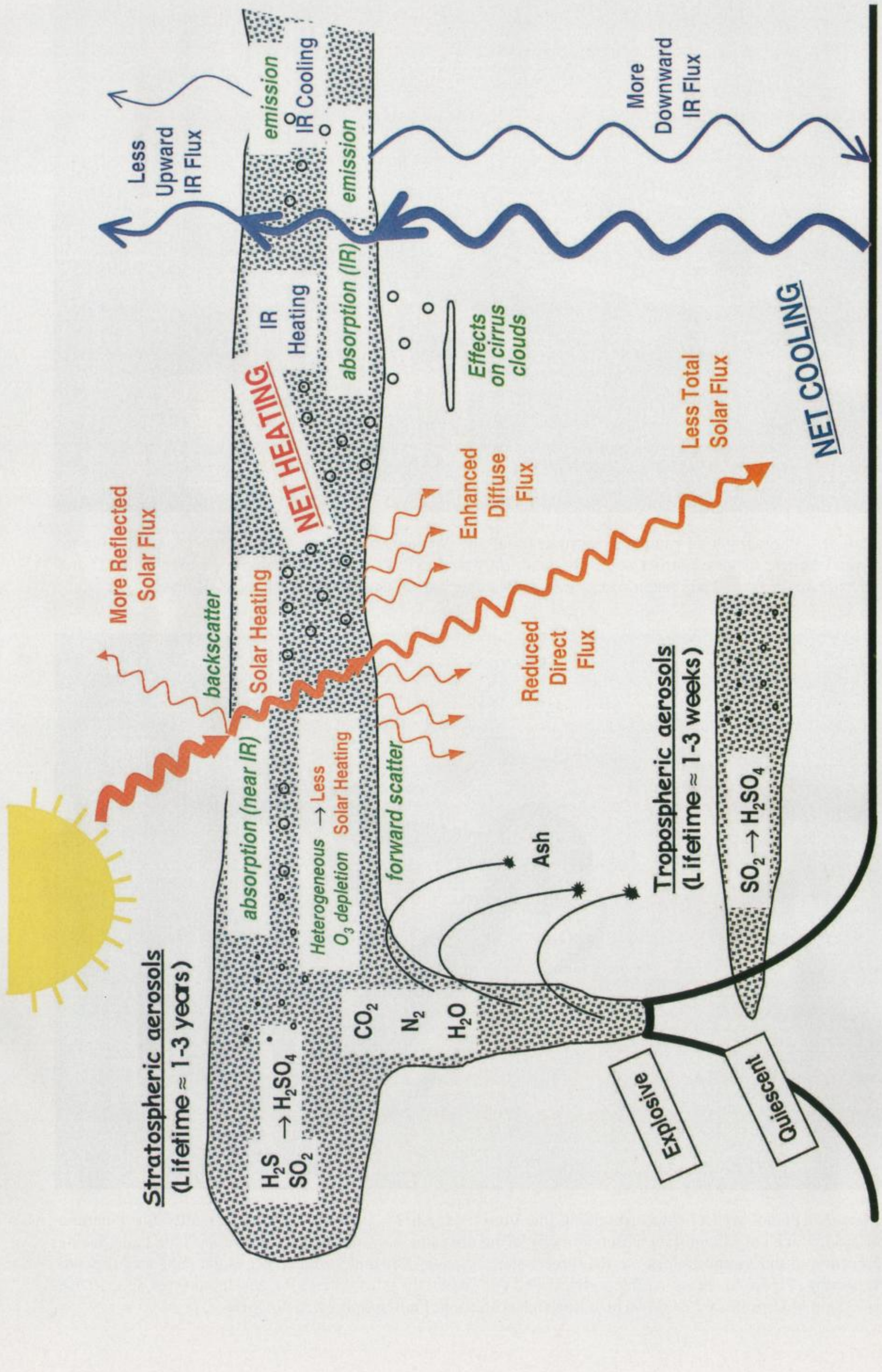


Plate 1. Schematic diagram of volcanic inputs to the atmosphere and their effects. This is an extended version of Figures 1 and 2 of *Simarski* [1992], drawn by L. Walter and R. Turco.

Plate 2. Photograph of radiation instruments at the Mauna Loa Observatory on the Island of Hawaii, United States, looking north toward Hualalai, on March 27, 1992. Observations in Figures 1 and 2 and photograph in Plate 3 are taken from these and similar instruments. Photograph by A. Robock.

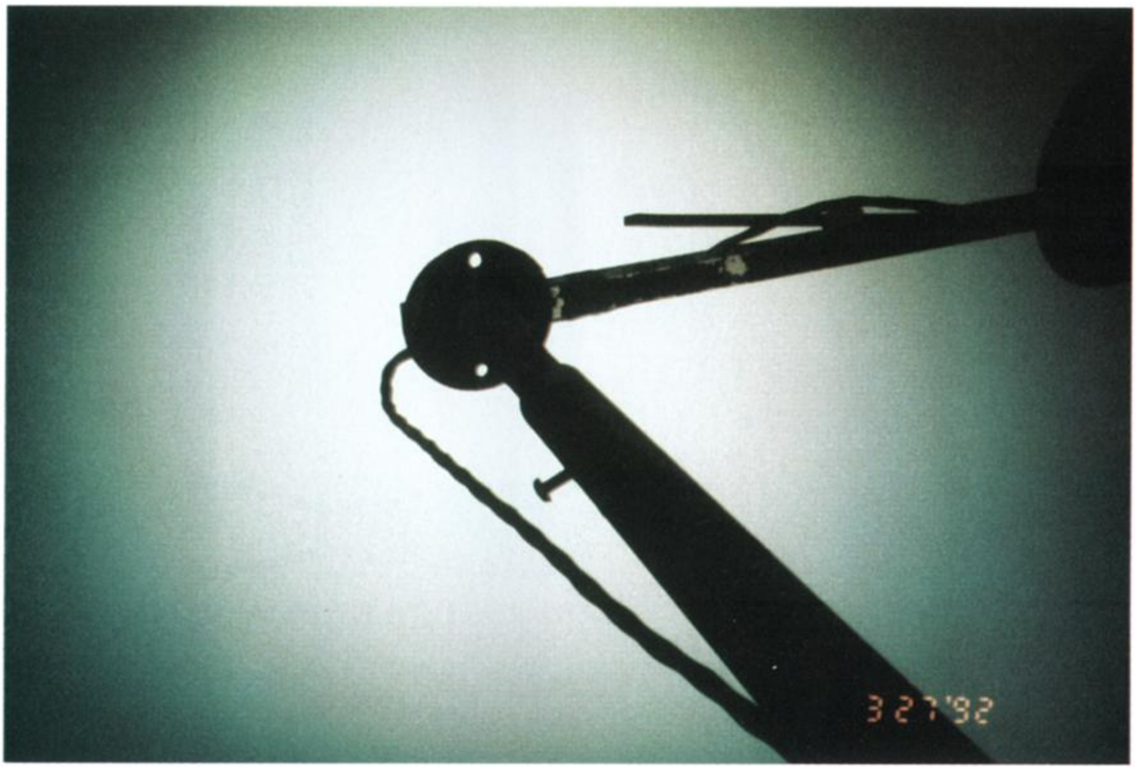


Plate 3. Photograph of sky surrounding the Sun on March 27, 1992, less than 1 year after the Pinatubo eruption, taken at Mauna Loa Observatory by laying the camera on the support shown in Plate 2 and having a portion of that support block out the direct solar radiation. The milky appearance is the enhanced forward scattering (Figure 2), clearly visible to the naked eye. When the stratosphere is clear, the normal appearance is a deep blue produced by molecular Rayleigh scattering. Photograph by A. Robock.

Plate 4. Sunset over Lake Mendota in Madison, Wisconsin, in May 1983, one year after the El Chichón eruption. Photograph by A. Robock.

easily be seen by the naked eye, making the normally blue sky milky white (forward scattering effect) (Plate 3). Volcanic aerosol clouds are clearly visible from space in solar wavelength images [e.g., *Robock and Matson*, 1983] and in space shuttle photographs (backscattering). The reflection of the setting Sun from the bottom of stratospheric volcanic aerosol layers (called "dust veils" by *Lamb* [1970]) produces the characteristic red sunset (Plate 4) used by Lamb as one means of detecting past eruptions. The famous 1893 Edvard Munch painting, "The Scream," shows a red volcanic sunset over the Oslo harbor produced by the 1892 Awu eruption. The timing of the reports of red sunsets was used by *Symons* [1888] and *Lamb* [1970] to calculate the height of the aerosol layer, taking into account the geometry of the setting Sun. *Robock* [1983a] also provides a diagram showing how these red sunsets can be observed after the Sun has set.

To evaluate the effects of a volcanic eruption on climate, the radiative forcing from the aerosols must be calculated. *Stenchikov et al.* [1998] presented a detailed study of the radiative forcing from the 1991 Mount Pinatubo eruption. Although this was the most comprehensively observed large eruption ever [e.g., *Russell et al.*, 1996], they still needed to make several assumptions to compensate for gaps in the observations. For globally smooth radiative perturbations, such as changing greenhouse gas concentrations, radiative forcing is defined as the change in the net radiative flux at the tropopause [*Houghton et al.*, 1996, p. 109]. For aerosols with a nonuniform vertical and horizontal distribution, *Stenchikov et al.* [1998] showed that a complete formulation of radiative forcing must include not only the changes of net fluxes at the tropopause, but also the vertical distribution of atmospheric heating rates and the change of downward thermal and net solar radiative fluxes at the surface. Using a detailed data set they developed from satellite and ground-based observations, they calculated the aerosol radiative forcing with the ECHAM4 (European Center/Hamburg) general circulation model (GCM) [*Roeckner et al.*, 1996].

An example of the radiative heating from Pinatubo is shown in Plate 5. At the top of the aerosol cloud, the atmosphere is warmed by absorption of solar radiation in the near infrared (near-IR). This effect dominates over the enhanced IR cooling due to the enhanced emissivity because of the presence of aerosols. *Andronova et al.* [1999] recently repeated these calculations with a more detailed radiation model and confirmed the importance of near-IR abosorption. In the lower stratosphere the atmosphere is heated by absorption of upward longwave radiation from the troposphere and surface. Hence this warming would be affected by the distribution of clouds in the troposphere, but *Stenchikov et al.* [1998] found that this effect (on changed upward longwave flux) was random and an order of magnitude smaller than the amplitude of the warming. In the troposphere, there are small radiative effects, since the

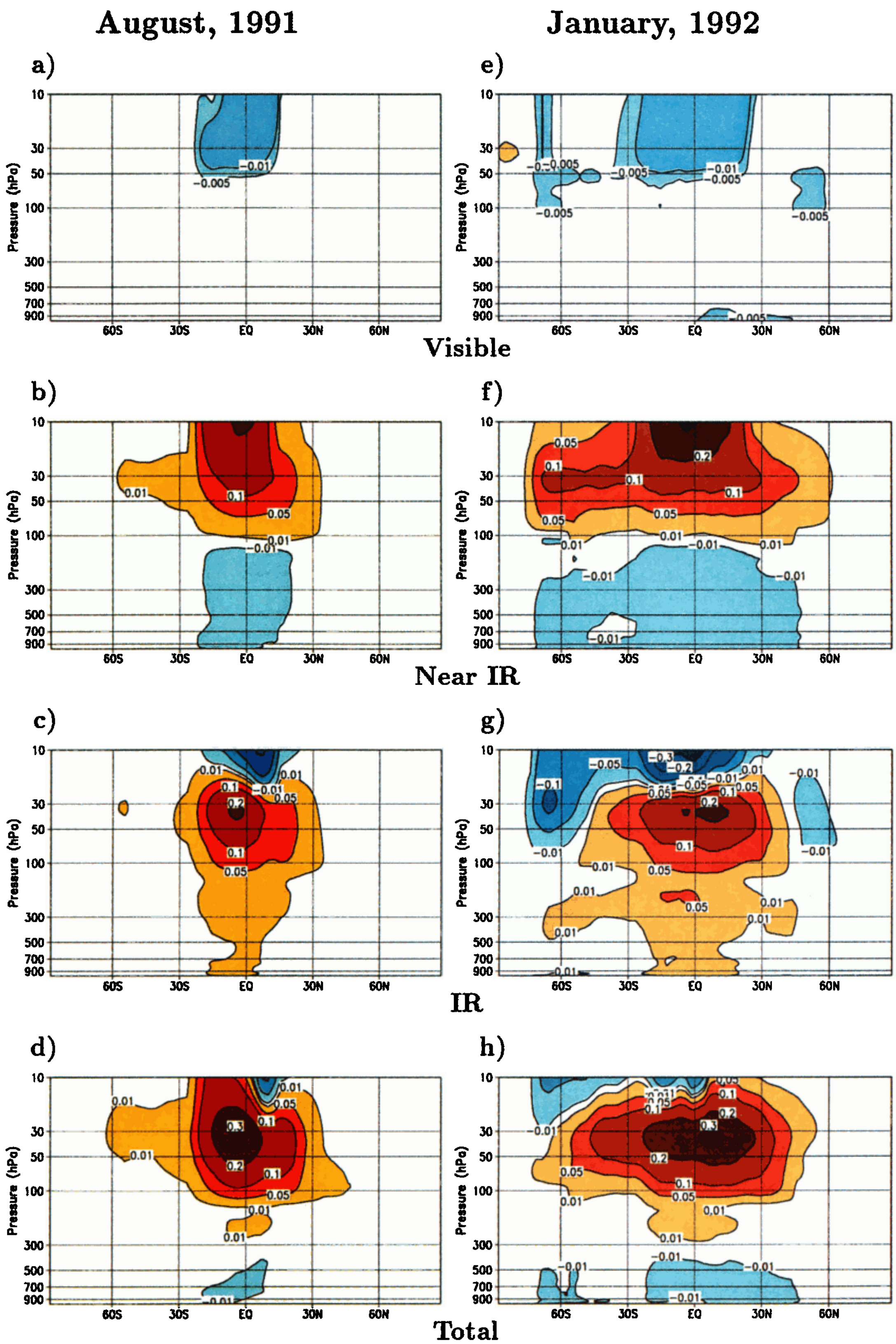


Plate 5. Radiative heating from Pinatubo for three different wavelength bands, from Figure 10 of *Stenchikov et al.* [1998]. Shown are monthly average, zonal average, perturbations of the radiative heating rates (K d^{-1}) caused by the Pinatubo aerosols for (a) visible ($\lambda \leq 0.68$ μm), (b) near-IR (0.68 μm $< \lambda <$ 4 μm), (c) IR ($\lambda \geq$ 4 μm), and (d) total, for August 1991, and (e) visible, (f) near-IR, (g) IR, and (h) total, for January 1992.

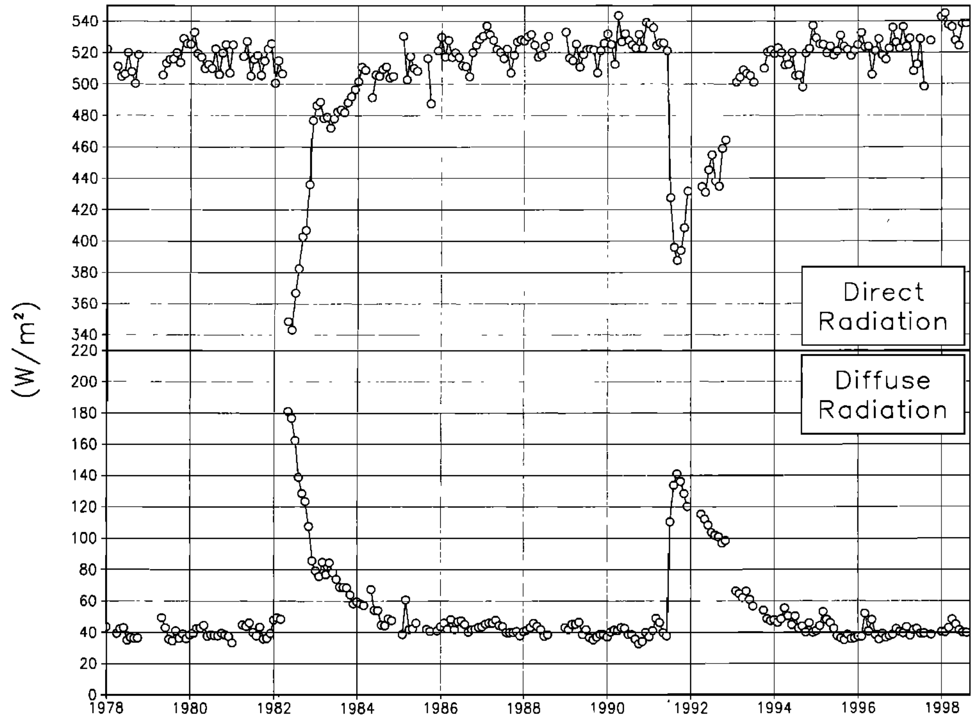


Figure 2. Direct and diffuse broadband radiation measurements from the Mauna Loa observatory, measured with a tracking pyrheliometer and shade disk pyranometer on mornings with clear skies at solar zenith angle of 60°, equivalent to two relative air masses [*Dutton and Christy*, 1992]. The reduction of direct radiation and enhancement of diffuse radiation after the 1982 El Chichón and 1991 Pinatubo eruptions are clearly seen. Years on abscissa indicate January of that year. Data courtesy of E. Dutton.

reduced downward near-IR (producing less absorption by water vapor in the troposphere) is compensated by the additional downward longwave radiation from the aerosol cloud. At the surface the large reduction in direct shortwave radiation overwhelms the additional downward diffuse shortwave flux and the small additional downward longwave radiation from the aerosol cloud, except in the polar night, where there is no sunlight, which was first shown by *Harshvardhan* [1979]. This net cooling at the surface is responsible for the well-known global cooling effect of volcanic eruptions. These calculations agree with observations of surface flux changes made at Mauna Loa by *Dutton and Christy* [1992]. They also agree with observations of *Minnis et al.* [1993] made with Earth Radiation Budget Experiment satellite data [*Barkstrom*, 1984], if the effects of stratospheric warming are considered.

Plate 5 also shows, for Pinatubo, that the lower stratospheric heating is much larger in the tropics than at the poles. It is this latitudinal gradient of heating which sets up a dynamical response in the atmosphere, resulting in the winter warming of NH continental regions, due to *advective effects*, which dominate over the radiative effects in the winter.

The possible effect of the aerosols on seeding cirrus cloud formation [*Mohnen*, 1990] is indicated in Plate 1. While evidence exists for individual cases of cirrus cloud formation by volcanic aerosols entering the troposphere through tropopause folds [*Sassen et al.*, 1995], the global effect has not been quantified.

4. INDICES OF PAST VOLCANISM

To evaluate the causes of climate change during the past century and a half of instrumental records or during the past 2000 years, including the *Medieval Warming* and the "Little Ice Age," a reliable record of the volcanic aerosol loading of the atmosphere is necessary. Five such indices (Table 2) have been compiled, based on different data sources and criteria, but none is perfect. *Robock and Free* [1995, 1996] describe these indices in detail and compare them, and here I summarize them. Another index described by *Robock and Free* [1995], being used by some climate modeling groups at the time, was never published and so is not included here. *Pollack et al.* [1976] also compiled a record of volcanic *optical depth*, but it was limited to 1880–1925 and 1962–1972.

A perfect index would convey the radiative forcing associated with each explosive eruption. The radiative forcing is most directly related to the sulfur content of emissions that reach into the stratosphere and not to the explosivity of the eruption. However, all indirect indices are either incomplete in geographical or temporal coverage or are a measure of some property of volcanic eruptions other than their stratospheric aerosol loading.

TABLE 2. Indices of Past Volcanic Eruptions

Name	*Units*	*How Calculated*	*Reference*
Dust veil index (DVI)	Krakatau = 1000	*Sapper* [1917, 1927], sunsets, eruption, and radiation observations	*Lamb* [1970, 1977, 1983]
Mitchell	aerosol mass	based on H. H. Lamb (personal communication, 1970)	*Mitchell* [1970]
Volcanic explosivity index (VEI)	Krakatau = 6	explosivity, from geologic and historical reports	*Newhall and Self* [1982] *Simkin et al.* [1981] *Simkin and Siebert* [1994]
Sato	τ (λ = 0.55 μm)	*Mitchell* [1970], radiation and satellite observations	*Sato et al.* [1993]
Ice core volcanic index (IVI)	τ (λ = 0.55 μm)	average of ice core acidity or sulfate measurements	*Robock and Free* [1995, 1996]

Direct radiation measurements would be the best technique, and combinations of surface, aircraft, balloon, and satellite measurements have clearly quantified the distributions and optical properties of the aerosols from the 1982 El Chichón [*Robock*, 1983a] and 1991 Pinatubo [*Stenchikov et al.*, 1998] eruptions (see also two special sections in *Geophysical Research Letters*: "Climatic Effects of the Eruption of El Chichón," *10*(11), 989–1060, 1983; and "The Stratospheric and Climatic Effects of the 1991 Mount Pinatubo Eruption: An Initial Assessment," *19*(2), 149–218, 1992). The brightness of the Moon during lunar eclipses can be used as an index of stratospheric turbidity, but such observations can only be made about once per year, sometimes missing the maximum aerosol loading [*Keen*, 1983]. Even these most recent large eruptions, however, have deficiencies in their observations [*Stenchikov et al.*, 1998]. In the past, however, such measurements are lacking, and indices have had to use the available surface radiation measurements combined with indirect measures such as reports of red sunsets in diaries and paintings, and geological evidence. Geological methods, based on examination of the deposits remaining on the ground from eruptions, can provide useful information on the total mass erupted and the date of the eruption, but estimates of the atmospheric sulfur loading are not very accurate. This "petrologic method" depends on the assumption that the difference in sulfur concentration between *glass inclusions* in the deposits near the volcano and the concentrations in the deposits themselves are representative of the total atmospheric sulfur injection, but this has been shown not to work well for recent eruptions for which we have atmospheric data [e.g., *Luhr et al.*, 1984].

For all the indices the problem of missing volcanoes and their associated dust veils becomes increasingly important the farther back in time they go. There may even have been significant volcanic aerosol loadings during the past century that do not appear in any or most of the volcanic indices. Volcanoes only appear in most of the indices if there is a report of the eruption from the ground. For recent eruptions, *Lamb* [1970] and *Sato et al.* [1993] used actual measurements of the radiative effects of the volcanic aerosols, and Lamb in addition used reports of atmospheric effects. Still, up to the present, all the indices may miss some Southern Hemisphere (SH) eruptions, as they may not be reported. Even in the 1980s, the December 1981 aerosols from the eruption of Nyamuragira were observed with lidar but were reported as the "mystery cloud" for several years until the source was identified by reexamining the TOMS satellite record [*Krueger et al.*, 1996]. As late as 1990, volcanic aerosols were observed with *Stratospheric Aerosol and Gas Experiment II* (*SAGE II*), but it has not been possible to identify the source [*Yue et al.*, 1994]. Before 1978, with no satellite or lidar records, there may be important missing eruptions even in the NH averages. This problem does not exist for individual ice core records, because they are objective measures of volcanic sulfuric acid, except that the farther back in time one goes with ice cores, the fewer such records exist, and each ice core record is extremely noisy and may have other problems. Plate 6 shows the five indices (Table 2) for the NH for the past 150 years. Here they are briefly described.

4.1. Lamb's Dust Veil Index

Lamb [1970, 1977, 1983] created a volcanic dust veil index (DVI), specifically designed for analyzing the effects of volcanoes on "surface weather, on lower and upper atmospheric temperatures, and on the large-scale wind circulation" [*Lamb*, 1970, p. 470]. *Lamb* [1970] and *Pollack et al.* [1976] suggested with data analyses that volcanism was an important cause of climate change for the past 500 years, and *Robock* [1979] used Lamb's index to force an energy-balance model simulation of the Little Ice Age, showing that volcanic aerosols played a major part in producing the cooling during that time period. The methods used to create the DVI, described by *Lamb* [1970] and *Kelly and Sear* [1982], include historical reports of eruptions, optical phenomena, radiation measurements (for the period 1883 onward), temperature information, and estimates of the volume of ejecta.

Lamb's DVI has been often criticized [e.g., *Bradley*, 1988] as having used climatic information in its derivation, thereby resulting in circular reasoning if the DVI is used as an index to compare with temperature changes. In fact, for only a few eruptions between 1763 and 1882 was the NH averaged DVI calculated based solely on temperature information, but for several in that period the DVI was calculated partially on the basis of temperature information. *Robock* [1981b] created a modified version of Lamb's DVI which excluded temperature information. When used to force a climate model, the results did not differ significantly from those using Lamb's original DVI, demonstrating that this is not a serious problem.

4.2. Mitchell Index

Mitchell [1970] also produced a time series of volcanic eruptions for the period 1850–1968 using data from Lamb. As discussed by *Robock* [1978, 1981b] and *Sato et al.* [1993], the Mitchell volcanic compilation for the NH is more detailed than Lamb's, because Lamb excluded all volcanoes with DVI < 100 in producing his NH annual average DVI. Mitchell provided a table of the order of magnitude of total mass ejected from each volcano, which is a classification similar to the volcanic explosivity index.

4.3. Volcanic Explosivity Index

A comprehensive survey of past volcanic eruptions [*Simkin et al.*, 1981; *Simkin and Siebert*, 1994] produced a tabulation of the volcanic explosivity index (VEI) [*Newhall and Self*, 1982] for all known eruptions, which gives a geologically based measure of the power of the volcanic explosion. This index has been used without any modification in many studies [see *Robock*, 1991] as an index of the climatological impact of volcanoes. A careful reading of *Newhall and Self* [1982], however, will find the following quotes: "We have restricted ourselves to consideration of volcanological data (no atmospheric data)..." (p. 1234) and "Since the abundance of sulfate aerosol is important in climate problems, VEI's must be combined with a compositional factor before use in such studies" (pp. 1234–1235). In their Table 1, Newhall and Self list criteria for estimating the VEI in "decreasing order of reliability," and the very last criterion out of 11 is "stratospheric injection." For VEI of 3, stratospheric injection is listed as "possible," for 4 it is "definite," and for 5 and larger it is "significant." If one attempts to work backward and use a geologically determined VEI to give a measure of stratospheric injection, serious errors can result. Not only is stratospheric injection the least reliable criterion for assigning a VEI, but it was never intended as a description of the eruption which had a VEI assigned from more reliable evidence. Nevertheless, *Robock and Free* [1995] found the VEI positively correlated with other indices, but imperfectly. For example, the Mount St. Helens eruption of 1980 has a large VEI of 5, and while it had a large local temperature impact [*Robock and Mass*, 1982; *Mass and Robock*, 1982], it had a negligible stratospheric impact [*Robock*, 1981a].

4.4. Sato Index

Sato et al. [1993] produced monthly NH and SH average indices. Their index, expressed as optical depth at wavelength 0.55 μm, is based on volcanological information about the volume of ejecta from *Mitchell* [1970] from 1850 to 1882, on optical extinction data after 1882, and on satellite data starting in 1979. The seasonal and latitudinal distributions for the beginning of the record are uniform and offer no advantages over the DVI and in fact show less detail than the latitudinally dependent index of *Robock* [1981b], who distributed the aerosols in latitude with a simple diffusive model. The more recent part of the record would presumably be more accurate than the DVI or VEI, as it includes actual observations of the latitudinal and temporal extent of the aerosol clouds.

4.5. Ice Core Volcanic Index

Robock and Free [1995] examined eight NH and six SH ice core records of acidity or sulfate for the period 1850 to the present in an attempt to identify the volcanic signal common to all records. They explain in detail the possible problems with using these records as volcanic indices, including other sources of acids and bases, other sources of sulfate, dating, local volcanoes, variable atmospheric circulation, the stochastic nature of snowfall and dry deposition, mixing due to blowing snow, and uncertainties in the electrical conductivity measures of the ice. For the NH, although the individual ice core records are, in general, not well correlated with each other or with any of the indices, a composite derived from averaging the cores, the ice core volcano index (IVI), showed promise as a new index of volcanic aerosol loading. This new index correlated well with the existing non-ice-core volcanic indices and with high-frequency temperature records. Still, it is clear that high-latitude volcanoes are given too much weight, and it is only possible to adjust for them if their signals can unambiguously be identified. For the SH the individual ice cores and indices were better correlated. The SH IVI was again highly correlated with all indices and individual ice cores but not with high-frequency temperature records.

Robock and Free [1996] attempted to extend the IVI farther into the past. They compared all the ice cores available for the past 2000 years with the DVI and the VEI for this period. An IVI constructed for the period 453 A.D. to the present showed little agreement with the DVI or VEI. They determined that except for a very few eruptions, the ice core record currently available is insufficient to delineate the climatic forcing by explosive volcanic eruptions before about 1200 for the NH and before about 1850 for the SH. They also point out, however, that the record of past volcanism remains buried in the ice of Greenland and Antarctica, and more

TABLE 3. Effects of Large Explosive Volcanic Eruptions on Weather and Climate

Effect	*Mechanism*	*Begins*	*Duration*
Reduction of diurnal cycle	blockage of shortwave and emission of longwave radiation	immediately	1–4 days
Reduced tropical precipitation	blockage of shortwave radiation, reduced evaporation	1–3 months	3–6 months
Summer cooling of NH tropics and subtropics	blockage of shortwave radiation	1–3 months	1–2 years
Stratospheric warming	stratospheric absorption of shortwave and longwave radiation	1–3 months	1–2 years
Winter warming of NH continents	stratospheric absorption of shortwave and longwave radiation, dynamics	$\frac{1}{2}$ year	one or two winters
Global cooling	blockage of shortwave radiation	immediately	1–3 years
Global cooling from multiple eruptions	blockage of shortwave radiation	immediately	10–100 years
Ozone depletion, enhanced UV	dilution, heterogeneous chemistry on aerosols	1 day	1–2 years

deep cores that analyze the sulfur or acid content have the promise of producing a reliable record of the past.

5. WEATHER AND CLIMATE RESPONSE

Volcanic eruptions can affect the climate system on many timescales (Table 3). The greatest known eruption of the past 100,000 years was the Toba eruption of about 71,000 years B.P. [*Zielinski et al.*, 1996], which occurred intriguingly close to the beginning of a major glaciation, and while *Rampino and Self* [1992] suggested a cause and effect relationship, it has yet to be established [*Li and Berger*, 1997]. Many papers, as discussed in the introduction, have suggested that volcanic aerosols can be important causes of temperature changes for several years following large eruptions and that even on a 100-year timescale, they can be important when their cumulative effects are taken into account. The effects of volcanic eruptions on climate are very significant in analyzing the global warming problem, as the impacts of anthropogenic greenhouse gases and aerosols on climate must be evaluated against a background of continued natural forcing of the climate system from volcanic eruptions, solar variations, and internal random variations from land-atmosphere and ocean-atmosphere interactions.

Individual large eruptions certainly produce global or hemispheric cooling for 2 or 3 years [*Robock and Mao*, 1995], and this signal is now clearer. The winter following a large tropical eruption is warmer over the NH continents, and this counterintuitive effect is due to nonlinear response through atmospheric dynamics [*Robock and Mao*, 1992; *Graf et al.*, 1993; *Mao and Robock*, 1998; *Kirchner et al.*, 1999]. Volcanic aerosols provide a surface for heterogeneous chemical reactions that destroy ozone, and observations following Pinatubo have documented midlatitude ozone depletion caused by a volcanic eruption [*Solomon et al.*, 1996; *Solomon*, 1999]. While the large 1982–1983 El Niño amplified just after the 1982 El Chichón eruption in Mexico, there is no evidence of a cause and effect relationship for this or any other eruptions [*Robock et al.*, 1995; *Robock and Free*, 1995; *Self et al.*, 1997]. Volcanic eruptions can still have a large local effect on surface temperatures in regions near the eruption for several days, as *Robock and Mass* [1982] and *Mass and Robock* [1982] showed for the 1980 Mount St. Helens eruption. In this section I briefly summarize these climatic effects, starting from the shortest timescale.

5.1. Reduction of Diurnal Cycle

The Mount St. Helens eruption in May 1980, in Washington State in northwestern United States, was a very powerful lateral blast which produced a huge local tropospheric loading of volcanic ash. In Yakima, Washington, 135 km to the east, it was so dark that automatic streetlights went on in the middle of the day. This thick aerosol layer effectively radiatively isolated the Earth's surface from the top of the atmosphere. The surface air temperature in Yakima was 15°C for 15 straight hours, independent of the normal diurnal cycle (Figure 3). *Robock and Mass* [1982] and *Mass and Robock* [1982] examined the errors of the model output statistics (MOS) forecasts produced by the National Weather Service. As the MOS forecasts did not include volcanic aerosols as predictors, they were able to interpret these errors as the volcanic effect. They found that the aerosols cooled the surface by as much as 8°C in the daytime but warmed the surface by as much as 8°C at night.

The reduction of the diurnal cycle only lasted a couple of days, until the aerosol cloud dispersed. The effect was also observed after the Krakatau eruption in Batavia (now know as Jakarta), Indonesia [see *Simkin and Fiske*, 1983, Figure 58]. While the Mount St. Helens eruption had a large local effect on temperature, no other impact was identified on precipitation or atmospheric circulation. Its stratospheric input of sulfur was very small, and hence this very explosive eruption had a minimal impact on global climate [*Robock*, 1981a].

5.2. Summer Cooling

It has long been known that the global average temperature falls after a large explosive volcanic eruption

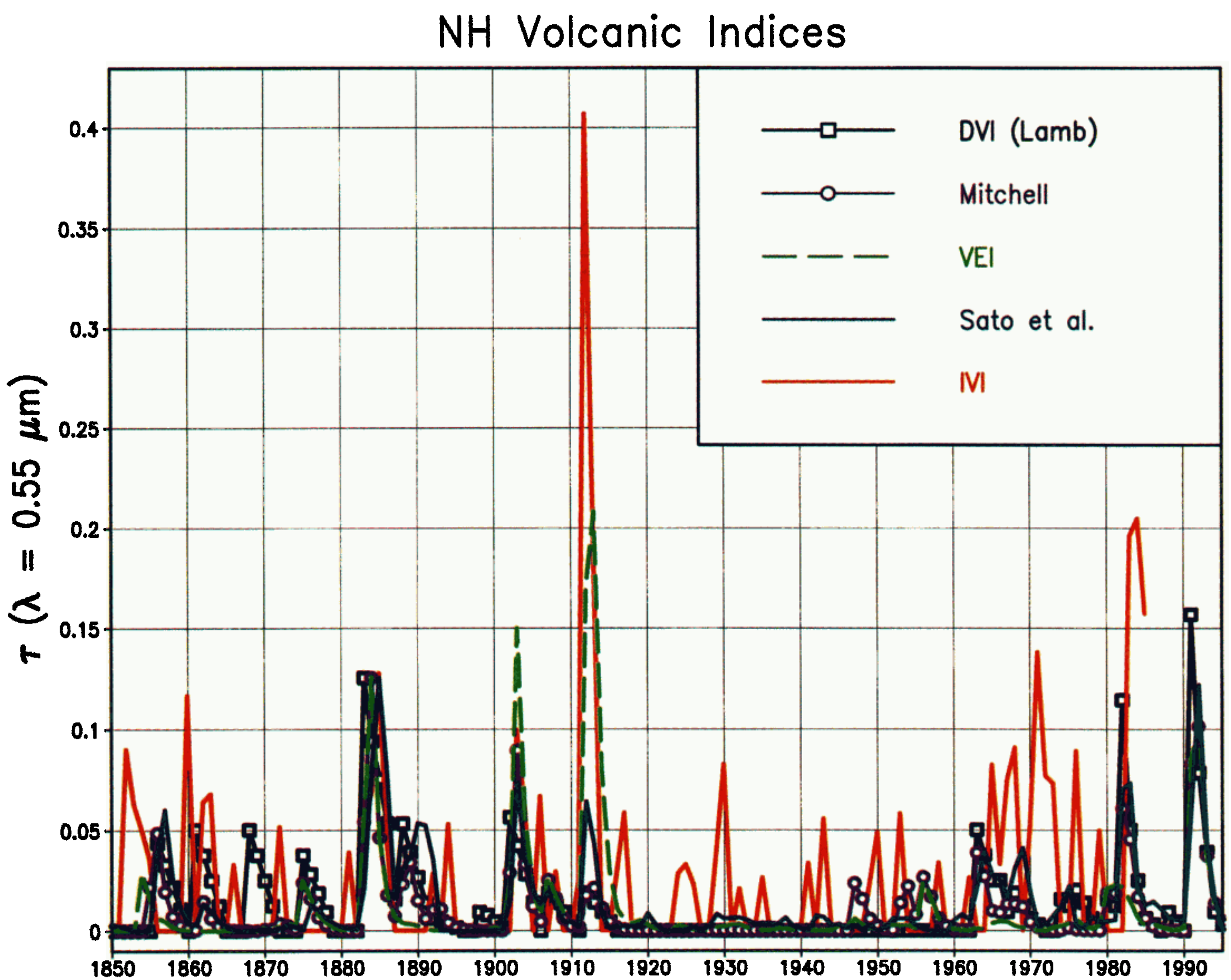


Plate 6. Graph of five different volcanic indices for the Northern Hemisphere, for 1850 to the present, from Figure 9 of *Robock and Free* [1995]. The indices are listed in Table 2 and described in the text. Here they are expressed as optical depth (τ) at λ = 0.55 μm, normalized to the values of *Sato et al.* [1993] for the Krakatau peak in the mid-1880s.

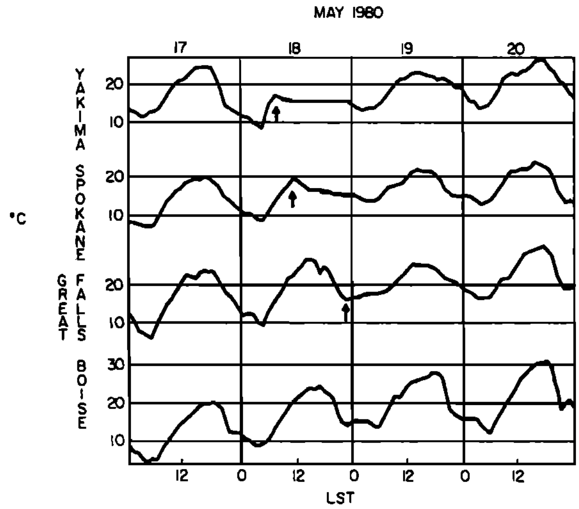


Figure 3. Time series of surface air temperature for Yakima and Spokane, Washington; Great Falls, Montana; and Boise, Idaho, for May 17–20, 1980, under the plume of the 1980 Mount St. Helens eruption, from Figure 3 of *Robock and Mass* [1982]. Time of arrival of the plume is indicated with an arrow. LST is local standard time. The plume never passed over Boise, which is included as a control. Note the damping of the diurnal cycle after the arrival of the tropospheric aerosol cloud.

[e.g., *Humphreys*, 1913, 1940; *Mitchell*, 1961]. The direct radiative forcing of the surface, with a reduction of total downward radiation, cools the surface. For example, *Hansen et al.* [1978], using a radiative-convective climate model, successfully modeled the surface cooling and stratospheric warming after the 1963 Agung eruption. In the tropics and in the midlatitude summer these radiative effects are larger than most other climatic forcings, as there is more sunlight to block. In some locations in the tropics, however, even the effects of a large eruption like El Chichón can be overwhelmed by a large El Niño, as was the case in 1983. The radiative-convective model study by *Vupputuri and Blanchet* [1984] also simulated cooling at the surface and warming in the stratosphere. Energy-balance models [*Schneider and Mass*, 1975; *Oliver*, 1976; *Bryson and Dittberner*, 1976; *Miles and Gildersleeves*, 1978; *Robock*, 1978, 1979; *Gilliland*, 1982; *Gilliland and Schneider*, 1984] have all shown cooling effects for up to several years after major eruptions. An early zonally averaged dynamic climate model [*MacCracken and Luther*, 1984] and GCM studies by *Hunt* [1977], *Hansen et al.* [1988, 1992, 1996], *Rind et al.* [1992], and *Pollack et al.* [1993] also showed volcanic cooling effects.

The problem of detection and attribution of a volcanic signal in the past is the same as the problem of identifying a greenhouse signal in the past: identifying a unique fingerprint of the forcing and separating out the effects of other potential forcings. It is necessary to separate the volcanic signal from that of other simultaneous climatic variations because the climatic signal of volcanic eruptions is of approximately the same amplitude as that of El Niño–Southern Oscillation (ENSO) and there have been so few large eruptions in the past century. For several recent eruptions, *Angell* [1988], *Nicholls* [1988], and *Mass and Portman* [1989] demonstrated that the ENSO signal in the past climatic record partially obscures the detection of the volcanic signal on a hemispheric annual average basis for surface air temperature.

More recently, *Robock and Mao* [1995] removed the ENSO signal from the surface temperature record to extract more clearly the seasonal and spatial patterns of the volcanic signal in surface temperature records. This observed signal matches the general cooling patterns found from GCM simulations [*Hansen et al.*, 1988; *Robock and Liu*, 1994]. Robock and Mao found, by superposing the signals of Krakatau, Santa Maria, Katmai, Agung, El Chichón, and Pinatubo, that the maximum cooling is found approximately 1 year following the eruptions. This cooling, of 0.1°–0.2°C, follows the solar declination but is displaced toward the NH (Figure 4); the maximum cooling in NH winter is at about 10°N and in summer is at about 40°N. This pattern is because of the distribution of continents: Land surfaces respond more quickly to radiation perturbations and thus the NH is more sensitive to the radiation reduction from volcanic aerosols.

Robock and Liu [1994] analyzed the *Hansen et al.* [1988] GCM simulations and found reduced tropical precipitation for 1–2 years following large eruptions. The global hydrological cycle is fueled by evaporation, and the cooling after the eruptions produced this effect. They even found a reduction in Sahel precipitation that matched the observed enhancement of the Sahel drought following the El Chichón eruption, but this result needs further confirmation before it can be considered robust.

5.3. Stratospheric Heating

It has long been known that the stratosphere is heated after injection of volcanic aerosols [e.g., *Quiroz*, 1983; *Parker and Brownscombe*, 1983; *Angell*, 1997b]. As explained above, this heating is caused by absorption of both near-IR solar radiation at the top of the layer and terrestrial radiation at the bottom of the layer. Figure 5 and Plate 7 shows observations of lower stratospheric temperature for the past 20 years. Two strong signals can be seen. After the 1982 El Chichón eruption the globally averaged stratospheric temperature rose by about 1°C for about 2 years. After the 1991 Pinatubo eruption a warming of equal length but about twice the amplitude is clearly visible. These large warmings are superimposed on a downward trend in stratospheric temperature caused by ozone depletion and increased CO_2 [*Ramaswamy et al.*, 1996; *Vinnikov et al.*, 1996]. This cooling, at 10 times the rate of tropospheric warming for the past century, is a clear signal of anthropogenic impacts on the climate system.

Plate 7 shows the stratospheric temperatures separated by latitude bands. After the 1982 El Chichón and

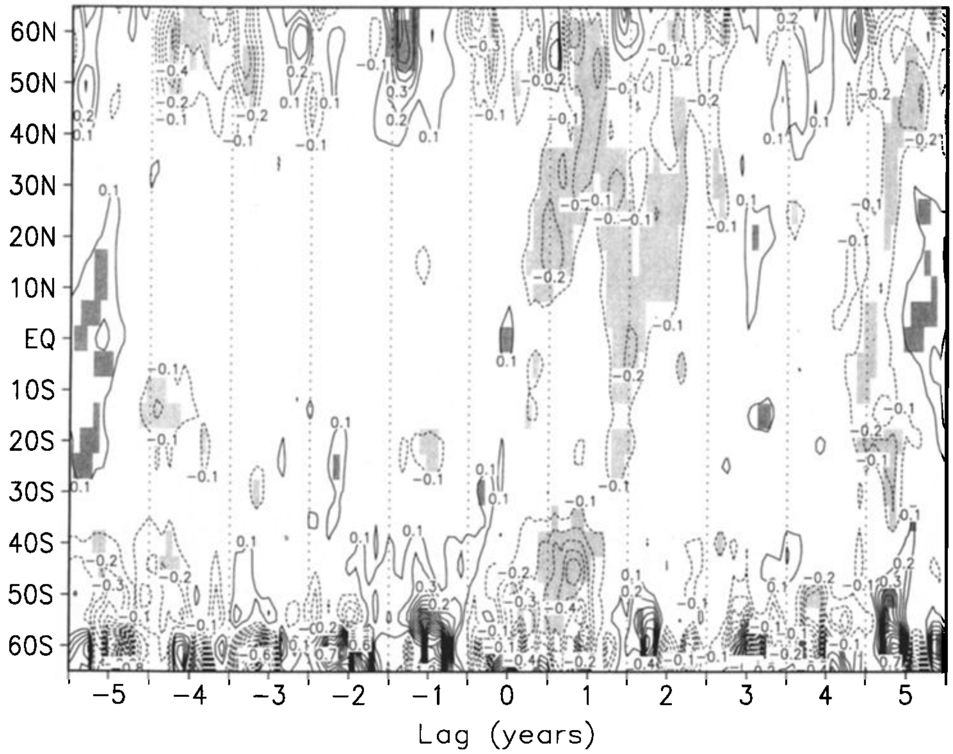


Figure 4. Zonal average surface air temperature anomalies, averaged for the six largest volcanic eruptions of the past century. Anomalies (°C) are with respect to the mean of the 5-year period before each eruption, with the seasonal cycle removed. Values significantly different (95% level) from 0°C are shaded. Contour interval is 0.1°C, with the 0°C interval left out and negative contours dashed. From Figure 6 of *Robock and Mao* [1995].

1991 Pinatubo eruptions the tropical bands (30°S–30°N), shown by the green and black curves, warmed more than the 30°N–90°N band (blue curve), producing an enhanced pole-to-equator temperature gradient. The resulting stronger polar vortex produces the tropospheric winter warming described next.

5.4. Winter Warming

Robock [1981b, 1984b] used an energy-balance climate model to examine the seasonal and latitudinal response of the climate system to the Mount St. Helens and El Chichón eruptions and found that the maximum surface cooling was in the winter in the polar regions of both hemispheres. This was due to the positive feedback of sea ice, which lowered the thermal inertia and enhanced the seasonal cycle of cooling and warming at the poles [*Robock*, 1983b]. Energy-balance models, however, are zonally averaged and parameterize atmospheric dynamics in terms of temperature gradients. Thus they do not allow nonlinear dynamical responses with zonal structure. While the sea ice/thermal inertia feedback is indeed part of the behavior of the climate system, we now know that an atmospheric dynamical response to large volcanic eruptions dominates the NH winter climate response, producing tropospheric warming rather than an enhanced cooling over NH continents.

5.4.1. Observations. It was first suggested by *Groisman* [1985], with reference to previous Russian studies, that warm winters over central Russia were a consequence of large volcanic eruptions. Using surface air temperature data for stations in Europe (including the European part of Russia) and northeastern North America for averages of two or three winters after the volcanic years of 1815, 1822, 1831, 1835, 1872, 1883, 1902, 1912, and 1963, he showed a significant warming over the central European part of Russia and insignificant changes in the other regions, including cooling over northeastern North America. In an update, *Groisman* [1992] examined the winter patterns after the 1982 El Chichón and 1991 Pinatubo eruptions and found warming over Russia again but found warming over northeastern North America.

By averaging over several winters and only examining part of the hemisphere, *Groisman* [1992] did not discover the complete winter warming pattern but did inspire other studies. His explanation of the warming over Russia as due to enhanced zonal winds bringing warm maritime air from the north Atlantic over the continent was correct, but he explained these enhanced winds as due to a larger pole-to-equator temperature gradient caused by polar cooling resulting from the eruptions. However, if the larger pole-to-equator temperature gra-

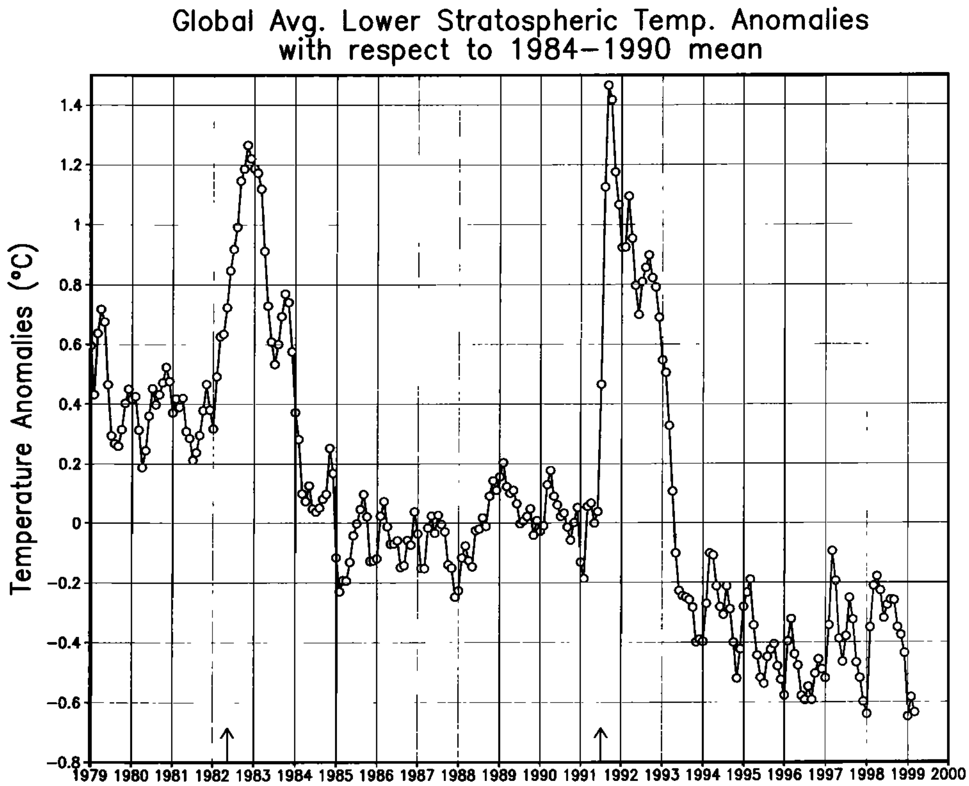
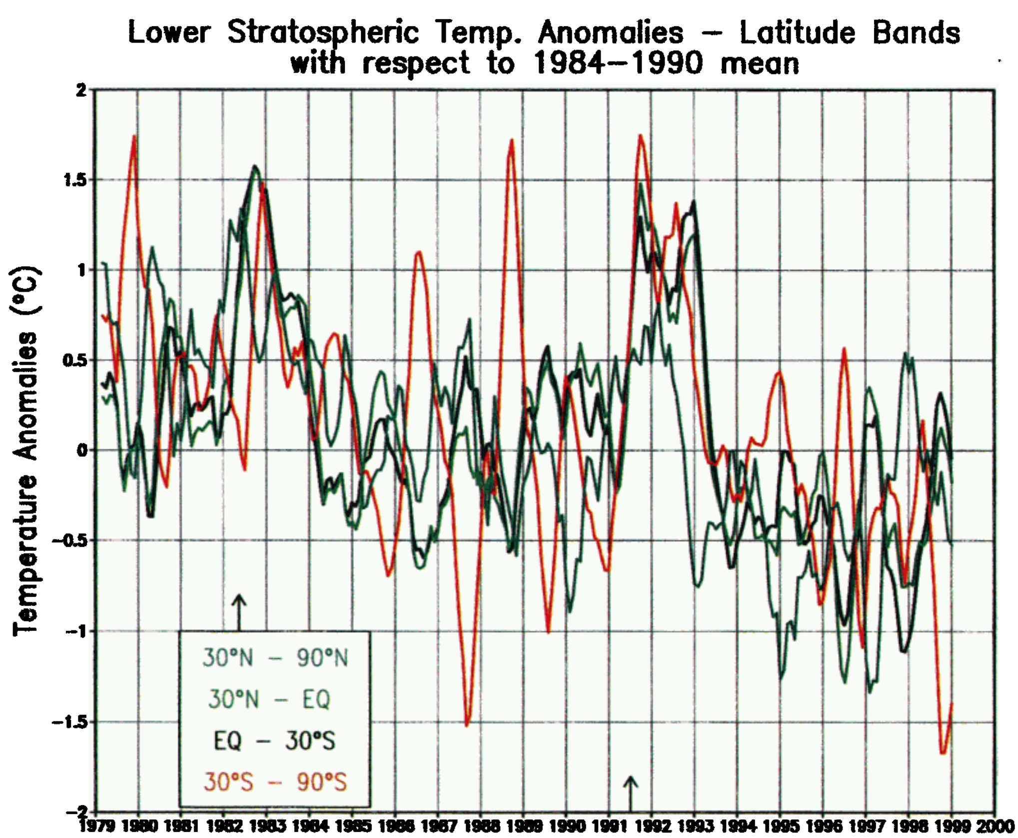


Figure 5. Global average monthly stratospheric temperatures from microwave sounding unit satellite observations, channel 4 [*Spencer et al.*, 1990] (updated in 1999). Anomalies (°C) are with respect to the 1984–1990 nonvolcanic period. Times of 1982 El Chichón and 1991 Pinatubo eruptions are denoted with arrows. Years on abscissa indicate January of that year.



dient was caused by polar cooling, why were the temperatures in the polar region over Eurasia warm and not cool? Why was northeastern North America cool after some eruptions and warm after others?

Lough and Fritts [1987], using tree-ring data and selecting 24 volcanic years between 1602 and 1900, found winter warming over western North America for the average of years 0–2 after the eruption years. The pattern they found is quite similar to that during ENSO years, and they made no attempt to correct for this factor. They also found spring and summer cooling in the central United States and summer warming over the United States west coast. They did not use any volcanoes south of 10°S, believing them not important for NH temperature variations.

Robock and Mao [1992] were the first to systematically examine the global surface air temperature record for the NH winter warming phenomenon. They examined the winter surface air temperature patterns after the 12 largest eruptions of the past century, removed the ENSO signal, and found a consistent pattern of warming over the continents and cooling over the oceans and the Middle East, when combining the first winter after tropical eruptions and the second winter after high-latitude eruptions. *Robock and Mao* [1995] showed that this pattern is mainly due to the tropical eruptions.

The winter warming pattern is illustrated in Plate 8, which shows the global lower tropospheric temperature anomaly pattern for the NH winter of 1991–1992, following the 1991 Mount Pinatubo eruption. This pattern is closely correlated with the surface air temperature pattern where the data overlap, but the satellite data allow global coverage. The temperature over North America, Europe, and Siberia was much warmer than normal, and that over Alaska, Greenland, the Middle East, and China was cold. In fact, it was so cold that winter that it snowed in Jerusalem, a very unusual occurrence. Coral at the bottom of the Red Sea died that winter [*Genin et al.*, 1995], because the water at the surface cooled and convectively mixed the entire depth of the water. The enhanced supply of nutrients produced anomalously large algal and phytoplankton blooms, which smothered the coral. This coral death had only happened before in winters following large volcanic eruptions [*Genin et al.*, 1995].

While the tropical regions cool in all seasons [*Robock and Mao*, 1995] (Figure 4), in the winter following a large eruption, the zonal average temperature change is small, but the wave pattern of anomalies produces large warm and cool anomalies, indicating a dynamical response in the climate system to the radiative forcing from the eruptions. The direct radiative response in the winter would be a small cooling, as it is winter because of the low insolation. At the pole the surface radiative forcing is actually warming, as there is a small increase in downward longwave radiation.

5.4.2. Theory. The winter warming patterns described above are closely related to tropospheric and stratospheric circulation. Both *Perlwitz and Graf* [1995] and *Kodera et al.* [1996] examined the observations of NH winter stratospheric and tropospheric circulation for the past 40 years and found that the dominant mode of circulation of the stratosphere is a strong polar vortex (polar night jet), which occurs simultaneously with a *500-mbar pattern* with a low anomaly over Greenland and high anomalies over North America, Europe, and east Asia. The associated surface air temperature pattern is exactly that shown in Plate 8. Perlwitz and Graf call this pattern the "baroclinic mode." The same pattern had previously been identified as the North Atlantic Oscillation (NAO) [*Hurrell*, 1995, and references therein] and is now also called the Arctic Oscillation (AO) [*Thompson and Wallace*, 1998, 2000a, b].

Both studies also found a second mode with weak stratospheric anomalies but a strong midtropospheric pattern of wave anomalies generated in the tropical Pacific and propagating across North America. This second barotropic mode [*Perlwitz and Graf*, 1995] has previously been identified as the Pacific North America *teleconnection pattern* associated with ENSO [*Wallace and Gutzler*, 1981]. It is confined to the Western Hemisphere and only influences extratropical regions over North America. *Kitoh et al.* [1996] found the same patterns in a GCM experiment, with the baroclinic mode dominating and no relationship between the ENSO mode and stratospheric circulation.

The picture that emerges is one of a characteristic baroclinic mode, or NAO or AO circulation pattern, in the troposphere (Figure 6) that produces anomalously warm surface air temperatures over the continents in the NH winter. It is a natural mode of oscillation of the winter atmospheric circulation, so external stratospheric forcing apparently can push the system into this mode without too much trouble. As *Perlwitz and Graf* [1995] explain in detail, the theoretical explanation is that the strong polar vortex traps the vertically propagating planetary waves, which constructively interfere to produce this stationary wave pattern. When there is a strong polar night jet, it prevents sudden stratospheric warmings [*McIntyre*, 1982] later in the winter and perpetuates

Plate 7. (opposite) Stratospheric temperature anomalies for four equal-area zonal bands, shown as 5-month running means, from microwave sounding unit satellite observations, channel 4 [*Spencer et al.*, 1990] (updated in 1999). Anomalies (°C) are with respect to the 1984–1990 nonvolcanic period. Times of 1982 El Chichón and 1991 Pinatubo eruptions are denoted with arrows. Note that after these eruptions the tropical bands warmed more than the 30°–90°N band, producing an enhanced pole-to-equator temperature gradient. Years on abscissa indicate January of that year.

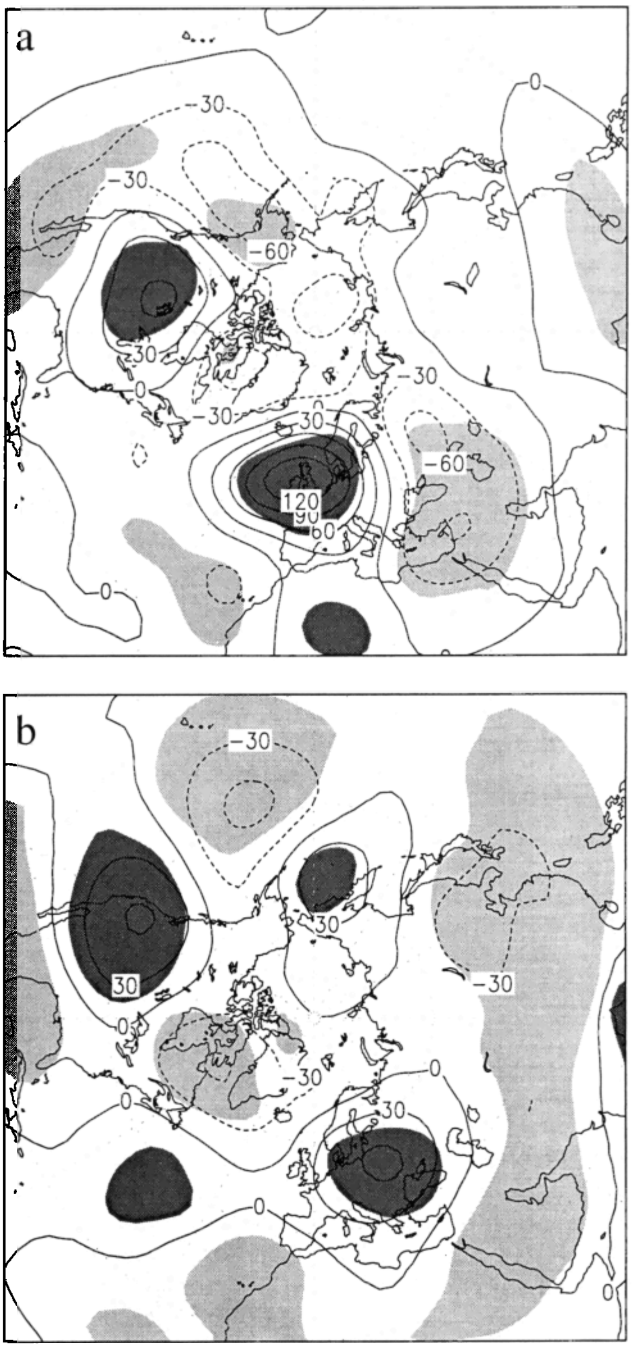


Figure 6. (a) Observed 500-mbar geopotential height (meters) anomaly pattern for the 1991–1992 NH winter (DJF) following the 1991 Mount Pinatubo eruption. Data are from the National Centers for Environmental Prediction reanalysis [*Kalnay et al.*, 1996], and anomalies are with respect to the period 1986–1990. (b) General circulation model (GCM) simulations from *Kirchner et al.* [1999]. Shown is the difference between the two GCM ensembles with and without stratospheric aerosols. In both panels, regions significantly different from 0 at the 80% level are shaded, with positive values shaded lighter and negative values shaded darker. The simulations reproduced the observed pattern of circulation anomalies, but not perfectly.

itself throughout the winter, keeping the polar lower stratosphere cold in the isolated vortex center. When the pattern occurs naturally without external forcing, it more frequently breaks down due to these sudden stratospheric warmings.

The winter warming circulation response occurs after all large tropical eruptions that have occurred since radiosonde observations began, Agung in 1963, El Chichón in 1982, and Pinatubo in 1991 [*Kodera*, 1994]. It should be noted, however, that other forcings can also work in the same direction. One is the 11-year solar cycle, as the 11-year cycle in ultraviolet flux and ozone amount combine to produce anomalous stratospheric heating and circulation [*Kodera et al.*, 1991; *Robock*, 1996a; *Haigh*, 1996; *Shindell et al.*, 1999]. These circulation changes help to explain the solar cycle–climate relations discovered by *Labitzke and van Loon* [1988] and *van Loon and Labitzke* [1990], but their patterns are still not completely understood. Another forcing is simply global warming, which increases the thickness of the tropical troposphere more than at higher latitudes, also enhancing the pole-to-equator gradient [*Graf et al.*, 1995]. Ozone depletion in the lower stratosphere has produced more cooling at the poles [*Ramaswamy et al.*, 1996], also enhancing the pole-to-equator gradient and producing a trend in the strength of the polar vortex [*Graf et al.*, 1995]. These global warming and ozone depletion trends help explain the observed trend in the NAO [*Hurrell*, 1995; *Hurrell and van Loon*, 1997] and the AO [*Thompson and Wallace*, 1998, 2000b] of the past several decades and in the related cold ocean–warm land pattern of *Wallace et al.* [1995, 1996]. Nevertheless, after a large tropical volcanic eruption, the forcing is so large that it overwhelms these more subtle trends and dominates the NH winter circulation of the succeeding year.

The connection between stratospheric circulation and surface air temperature anomalies in the NH winter after a large tropical volcanic eruption is illustrated schematically in Plate 9. The heating of the tropical lower stratosphere by absorption of terrestrial and solar near-IR radiation (Plate 5) expands that layer and produces and enhanced pole-to-equator temperature difference. The strengthened polar vortex traps the wave energy of the tropospheric circulation, and the stationary wave pattern known as the NAO or AO dominates the winter circulation, producing the winter warming.

5.4.3. Modeling. To test the above theory, *Graf et al.* [1993] presented results from a perpetual January GCM calculation of the effects of stratospheric aerosols on climate that showed winter warming over both northern Eurasia and Canada and cooling over the Middle East, northern Brazil, and the United States. They used forcing with a stratospheric aerosol loading in the pattern of the El Chichón volcano and the low-resolution (T21) ECHAM2 GCM, and the circulation response in January following volcanic eruptions was well simulated [*Kirchner and Graf*, 1995]. *Kirchner et al.* [1999] repeated this calculation in more detail with an improved model, in an experiment with better defined aerosol parameters [*Stenchikov et al.*, 1998], interactive calculation of the aerosol radiative effects, and an improved GCM (ECHAM4) [*Roeckner et al.*, 1996], including the full seasonal cycle. They were successful in reproducing the observed circulation (Figure 6) and surface temperature patterns (Figure 7) after the 1991 Pinatubo eruption when using climatological sea surface temperatures. Although the simulated patterns did not exactly match the observations, the positive height anomalies over North America and Europe and negative anomalies over Baffin Island and the Middle East (Figure 6) are in the right location, which correspond to warm and cold anomalies over the same locations (Figure 7). The simulations were not as successful when using the actual observed sea surface temperatures, which included a moderate El Niño. Thus GCMs have not yet demonstrated an ability to successfully simulate the complete response of the climate system to surface and stratospheric forcing. This is an area of active current research, including the new Pinatubo Simulation Task of the GCM–Reality Intercomparison Project for the Stratosphere (GRIPS) [*Pawson et al.*, 2000].

Mao and Robock [1998] conducted another model test of the winter warming pattern. Patterns similar to that in Plate 8 appeared over North America in the winters following the 1982 El Chichón and 1991 Pinatubo eruptions, which were both during El Niños. This led some to suggest that the temperature patterns were produced by the Pacific North American teleconnection pattern connected with the El Niño. As both forcings were acting at the same time, observations alone cannot distinguish the cause of the pattern. Mao and Robock took advantage of the design of the Atmospheric Model Intercomparison Project (AMIP) [*Gates*, 1992], which was conducted for the period 1979–1988. Thirty different GCMs simulated the weather of this 10-year period, forced only with observed sea surface temperatures (SSTs), which included two El Niños, 1982–1983 and 1986–1987. No volcanic aerosol forcing was used in the simulations, so the results of the simulation should reflect ENSO patterns only. Mao and Robock found that the North American patterns simulated by the models were very similar for both El Niño winters. They were very close to the observed pattern for 1986–1987, a winter with only El Niño forcing and no volcanic aerosols, but did not resemble the observed pattern of 1982–1983 (Plate 8) at all. Therefore they concluded that the major warming over the North American and Eurasian continents in the 1982–1983 NH winter is not an ENSO-dominant mode, but rather a pattern associated with the enhanced stratospheric polar vortex by the larger equator-to-pole temperature gradient produced by volcanic sulfate aerosols in the stratosphere.

Franklin [1784] noted that the winter in Europe following the 1783 Lakagigar eruption was extremely cold rather than warm. Because Lakagigar was a high-latitude eruption, it could have produced high-latitude stratospheric heating during the winter of 1783–1984 and a reduced pole-to-equator temperature gradient. This theory suggests that the opposite phase of the pattern discussed above was produced, with a large negative NAO anomaly. This suggestion is now being tested with GCM experiments.

5.5. Little Ice Age

Since volcanic aerosols normally remain in the stratosphere no more than 2 or 3 years, with the possible exception of extremely large eruptions such as that of Toba, approximately 71,000 years ago [*Bekki et al.*, 1996], the radiative forcing from volcanoes is interannual rather than *interdecadal* in scale. A series of volcanic eruptions could, however, raise the mean optical depth significantly over a longer period and thereby give rise to a decadal-scale cooling. If a period of active volcanism ends for a significant interval, the adjustment of the climate system to no volcanic forcing could produce warming. This was the case for the 50 years from 1912 to 1963, when global climate warmed. Furthermore, it is possible that feedbacks involving ice and ocean, which act on longer timescales, could transform the short-term volcanic forcing into a longer-term effect. As a result, the possible role of volcanoes in decadal-scale climate change remains unclear. In particular, the current century is the warmest of the past five, with the previous four centuries earning the moniker of the Little Ice Age due to its coldness. This period has never been satisfactorily explained. What was the role of volcanism in this climate change?

Until recently, *Schneider and Mass* [1975] and *Robock* [1979] were the only models to use volcanic chronologies to investigate periods before the midnineteenth century. At the same time, they examined the hypothesis that solar constant variations linked to the sunspot cycle were also an important cause of climate change on this timescale. Schneider and Mass used a zero-dimensional energy-balance model and found agreement of several large-scale features of their simulations with climate records. They concluded that while volcanoes had a weak relationship to climate, the solar-climate effect was not proven. Robock used a latitudinally resolved energy-balance model with volcanic [*Mitchell*, 1970] and solar forcing (proportional to the envelope of the sunspot number) and found the volcanic forcing explained a much larger share of the temperature variability since 1620 than did the solar series. Both the Schneider and Mass and Robock models used a simple mixed-layer ocean.

Three new studies [*Crowley and Kim*, 1999; *Free and Robock*, 1999; *D'Arrigo et al.*, 1999] have now made use of new volcano chronologies, new solar constant reconstructions, and new reconstructions of climate change for the Little Ice Age to address this problem again. They used upwelling-diffusion energy-balance climate models to simulate the past 600 years with volcanic, solar, and anthropogenic forcings and compared the

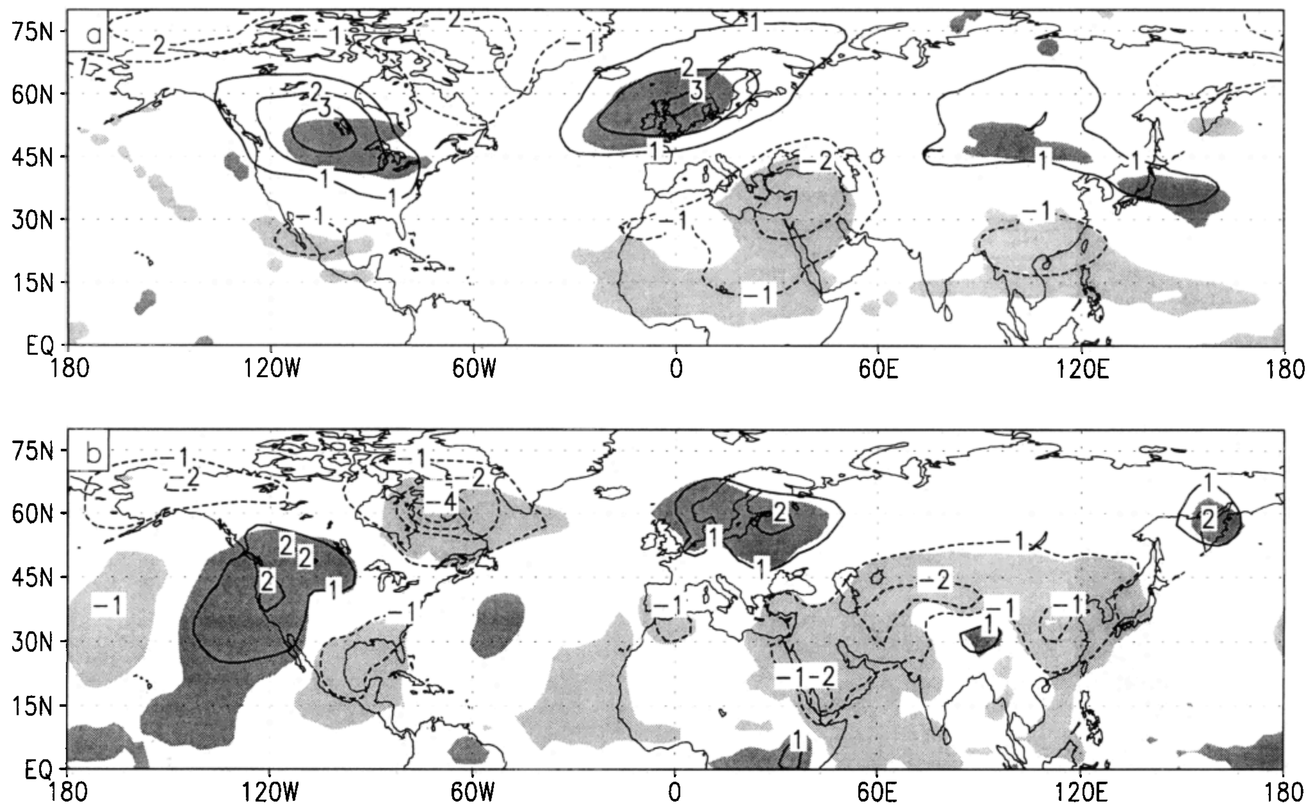


Figure 7. Temperature patterns for NH winter (DJF) 1991–1992 following the Mount Pinatubo eruption. (a) Observed anomalies (with respect to the average for 1986–1990) from channel 2LT of the microwave sounding unit satellite observations (same as in Plate 8), representative of the lower troposphere. (b) Anomalies simulated by a GCM forced with observed Pinatubo aerosols [*Kirchner et al.*, 1999] expressed as GCM simulated channel 2LT temperatures. Shown are differences between the two GCM ensembles with and without stratospheric aerosols. Contour interval is 1°C, and shading is as in Figure 6. The locations of anomalous warm and cold regions are quite close to those observed.

results with paleoclimatic reconstructions, mainly based on tree rings. They conclude, subject to the limitations of the forcing and validation data sets, that volcanic eruptions and solar variations were both important causes of climate change in the Little Ice Age. Furthermore, the warming of the past century cannot be explained by natural causes and can be explained by warming from anthropogenic greenhouse gases. Further work is needed in this area, however, linking ocean-atmosphere interactions with better volcanic and solar chronologies.

6. DO VOLCANIC ERUPTIONS PRODUCE EL NIÑOS?

The April 1982 El Chichón volcanic eruption was the largest of the century up to that time. Beginning in April 1982, as indicated by the Southern Oscillation Index, an unpredicted and unprecedented El Niño began, resulting in the largest warm ENSO event of the century up to that time, with record warm temperatures in the eastern equatorial Pacific Ocean and remote temperature and precipitation anomalies in distant locations [*Halpert and Ropelewski*, 1992]. Owing to the coincidence of the beginning of the ENSO event and the El Chichón eruption, several suggestions were made as to a cause and effect relationship, even going so far as to suggest that most El Niños were caused by volcanic eruptions.

How could a volcanic eruption produce an El Niño? Would the mechanism involve the stratospheric aerosol cloud, tropospheric aerosols, or the dynamical response to surface temperature changes? What is the evidence based on the past record of volcanic eruptions and ENSO events? Was the simultaneous occurrence of the El Chichón eruption and large El Niño just a coincidence, or was it an example of a mechanism that works after many other large eruptions?

Hirono [1988] proposed a plausible mechanism whereby tropospheric aerosols from the El Chichón eruption would induce an atmospheric dynamical response producing a trade wind collapse, and the trade wind reduction would produce an oceanographic response which affected the timing and strength of the resulting El Niño. He suggested that large tropospheric aerosols falling out of the El Chichón eruption cloud into the region west of Mexico in the east Pacific in the

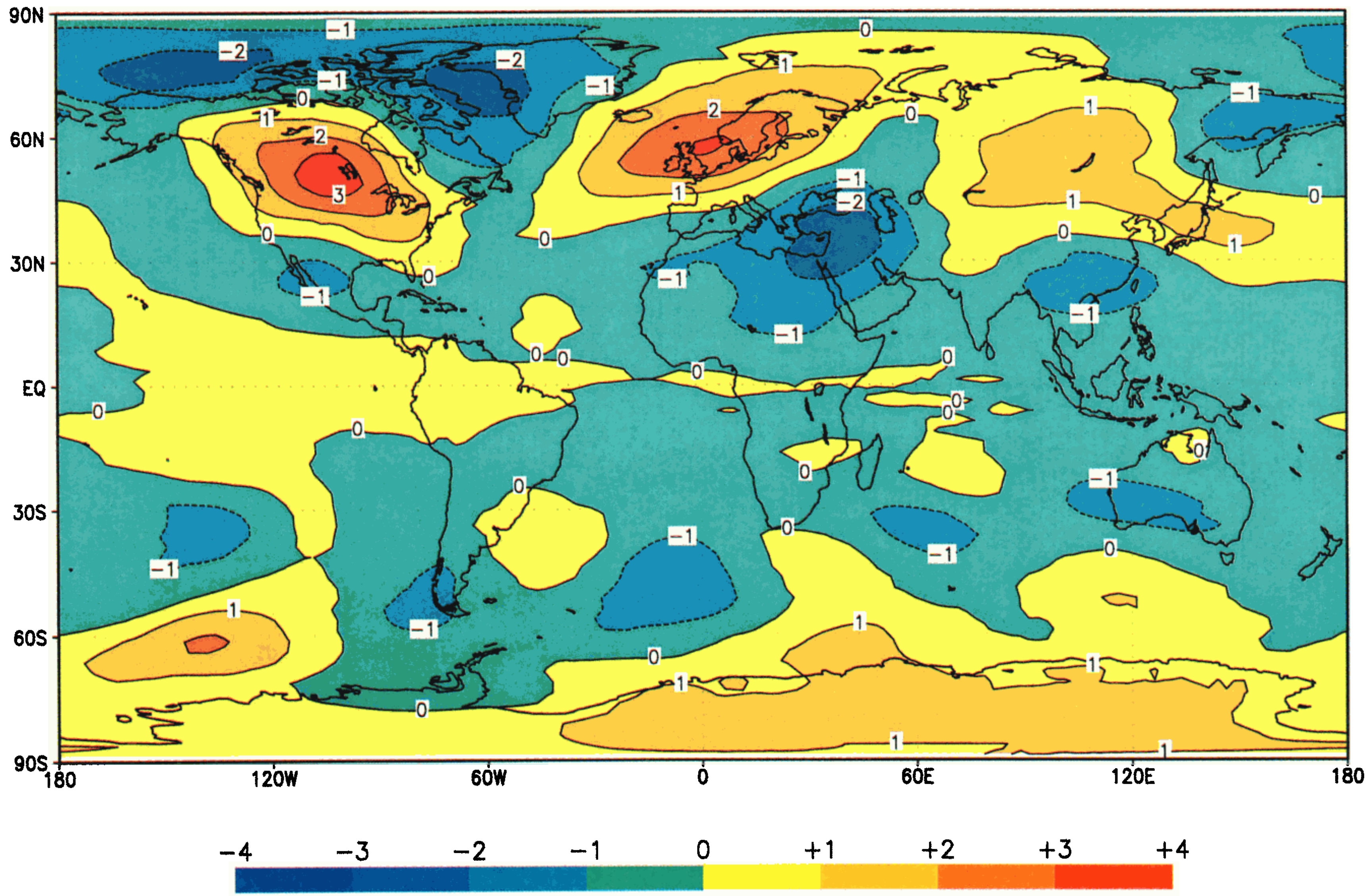


Plate 8. Winter (December-January-February (DJF)) lower tropospheric temperature anomalies (with the nonvolcanic period of 1984–1990 used to calculate the mean) for the 1991–1992 Northern Hemisphere (NH) winter following the 1991 Mount Pinatubo eruption. This pattern is typical of that following all large tropical eruptions, with warming over North America, Europe, and Siberia and cooling over Alaska, Greenland, the Middle East, and China. Data are from microwave sounding unit channel 2R [*Spencer et al.*, 1990], updated courtesy of J. Christy and now called channel 2LT.

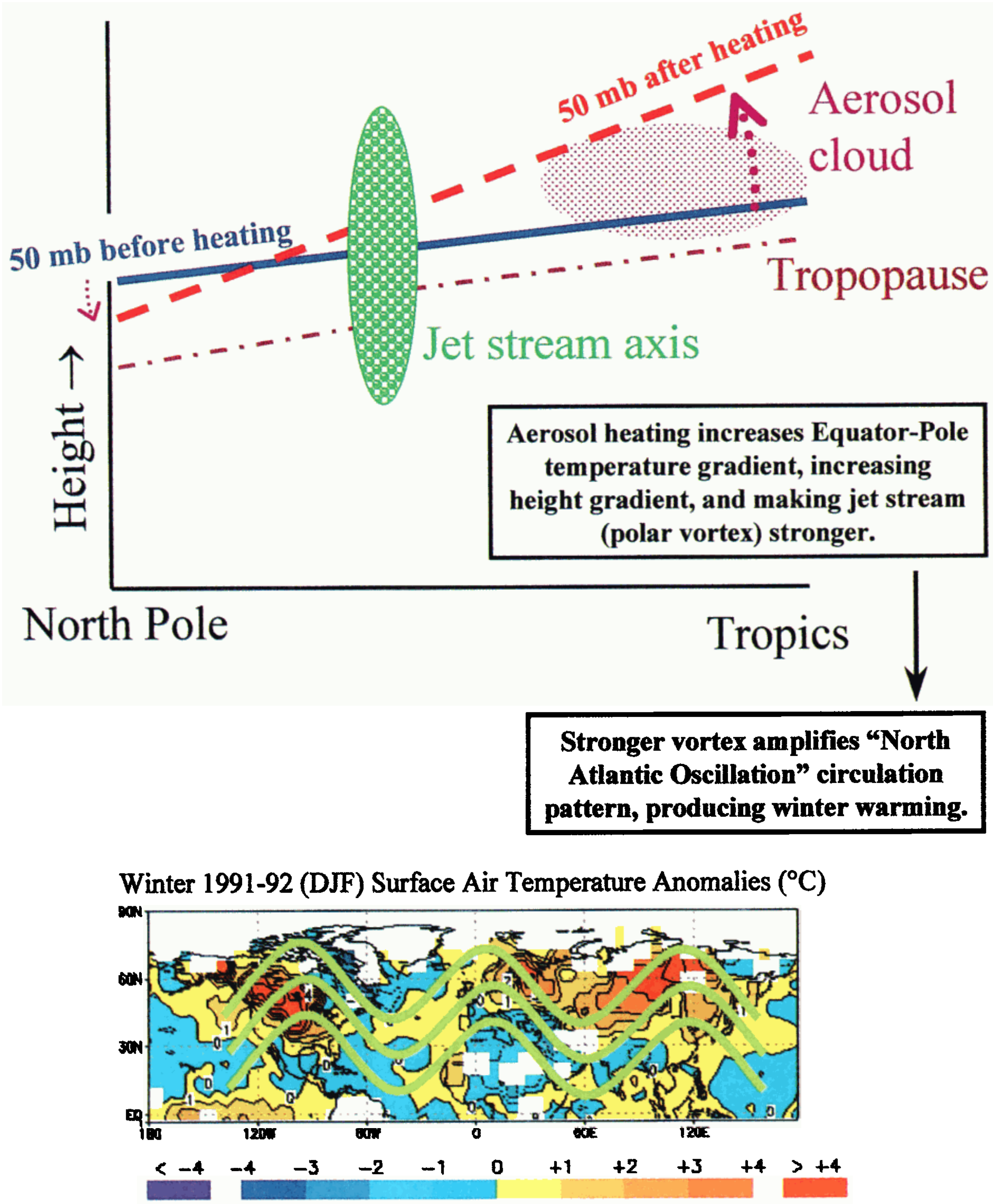


Plate 9. Schematic diagram of how tropical lower stratospheric heating from volcanic aerosols produces the winter warming temperature pattern at the surface. The map at the bottom is surface air temperature anomalies (with respect to 1961–1990) for the 1991–1992 NH winter (DJF) following the 1991 Mount Pinatubo eruption. The green curves indicate the anomalous tropospheric wind patterns responsible for horizontal advection that produced these temperature anomalies. This temperature pattern is very similar to the one shown in Plate 8 but is for the surface. Data courtesy of P. Jones.

weeks after the eruption would absorb terrestrial and solar radiation, heating the atmosphere and inducing a low-pressure region, changing the tropospheric circulation. The wind spiraling into the low would reduce the northeast trade winds, producing an oceanographic response that would cause an El Niño. *Robock et al.* [1995] used the Lawrence Livermore National Laboratory version of the Community Climate Model, Version 1 (CCM1) GCM to investigate this mechanism with radiative forcing from the observed aerosol distribution. Indeed, there was a trade wind collapse induced in the atmosphere, but it was too late and in the wrong position to have produced the observed El Niño.

It is interesting to note that when this study was beginning, E. Rasmusson, one of the world's experts on El Niños and whose office was next to mine at the University of Maryland, the week before the 1991 Pinatubo eruption excitedly walked down the hall announcing that a new El Niño was beginning. Thus although in 1991 there was again a large volcanic eruption at the same time as a large El Niño, the El Niño occurred first. In fact, *Rampino et al.* [1979] once suggested that climatic change could cause volcanic eruptions by changing the stress on the Earth's crust as the snow and ice loading shifts in response to the climate. More recent evidence of relatively large changes in the Earth's rotation rate after El Niños, in response to the enhanced atmospheric circulation [*Marcus et al.*, 1998], could produce an additional lithospheric stress. These speculations have not been proven, however, and further study is limited by available long-term records.

Schatten et al. [1984] and *Strong* [1986] suggested that the atmospheric dynamic response to the stratospheric heating from the volcanic aerosols would perturb the Hadley cell circulation in the tropics in such a way as to reduce the trade winds and trigger an El Niño. Schatten et al.'s model, however, was zonally symmetric and did not explain why the response would be so rapid and mainly in the Pacific.

Handler [1986] suggested that the climatic response to cooling over continents produces a monsoon-like circulation that somehow produces El Niños, but he did not clearly explain the physical mechanism or demonstrate that a model can produce this effect. His claim to find a statistical link between occurrence of volcanic eruptions and El Niños is flawed in several ways. As there have been many volcanic eruptions in the past few centuries and El Niños occur every 4–7 years, it is possible to find an eruption close in time to each El Niño. However, to establish a cause and effect relationship, the eruption must be of the proper type and at the proper time to produce the claimed response. Handler's studies did not use a proper index of stratospheric sulfate loading. He selected very small eruptions that could not have had a climatic effect, and he was not careful with the timing of the eruptions he chose, claiming eruptions that occurred after particular ENSO events or long before them were their causes.

Robock and Free [1995] calculated the correlation between the Southern Oscillation Index [*Ropelewski and Jones*, 1987], the best available index of past El Niños, and every volcanic index described above, as well as each individual ice core record that was available, and in no cases found significant correlations. *Self et al.* [1997] recently examined every large eruption of the past 150 years and showed that in no cases was there an El Niño that resulted as a consequence. Therefore, as concluded by *Robock et al.* [1995], the timing and location of the El Chichón eruption and the large ENSO event that followed were coincidental, and there is no evidence that large volcanic eruptions can produce El Niños.

7. EFFECTS ON STRATOSPHERIC OZONE

Volcanic aerosols have the potential to change not only the radiative flux in the stratosphere, but also its chemistry. The most important chemical changes in the stratosphere are related to O_3, which has significant effects on ultraviolet and longwave radiative fluxes. The reactions which produce and destroy O_3 depend on the UV flux, the temperature, and the presence of surfaces for heterogeneous reactions, all of which are changed by volcanic aerosols [*Crutzen*, 1976; *Tabazadeh and Turco*, 1993; *Tie and Brasseur*, 1995; *Tie et al.*, 1996; *Solomon et al.*, 1996]. The heterogeneous chemistry responsible for the ozone hole over Antarctica in October each year occurs on polar stratospheric clouds of water or nitric acid, which only occur in the extremely cold isolated spring vortex in the Southern Hemisphere. These reactions make anthropogenic chlorine available for chemical destruction of O_3. Sulfate aerosols produced by volcanic eruptions can also provide these surfaces at lower latitudes and at all times of the year.

Solomon [1999] describes the effects of aerosols on ozone in great detail. Here only a brief mention of some of the issues related to volcanic eruptions is made. Quantifying the effects of volcanic aerosols on ozone amount is difficult, as chemical and dynamical effects occur simultaneously and the effects are not much larger than natural variability. Nevertheless, attempts were made after the 1991 Pinatubo eruption to estimate the effects on ozone. Column O_3 reduction of about 5% was observed in midlatitudes [*Zerefos et al.*, 1994; *Coffey*, 1996], ranging from about 2% in the tropics to about 7% in the midlatitudes [*Angell*, 1997a]. Therefore ozone depletion in the aerosol cloud was much larger and reached about 20% [*Grant et al.*, 1992; *Grant*, 1996]. The chemical ozone destruction is less effective in the tropics, but lifting of low ozone concentration layers with the aerosol cloud [*Kinne et al.*, 1992] causes a fast decrease in ozone mixing ratio in the low latitudes. Similarly, subsidence at high latitudes increases ozone concentration there and masks chemical destruction.

Decrease of the ozone concentration causes less UV absorption in the stratosphere, which modifies the aerosol heating effect [*Kinne et al.*, 1992; *Rosenfield et al.*, 1997]. The net effect of volcanic aerosols is to increase surface UV [*Vogelmann et al.*, 1992]. The subsequent O_3 depletion allows through more UV than is backscattered by the aerosols.

Kirchner et al. [1999] calculated a January 1992 global average 70-mbar heating of about 2.7 K from the aerosol heating following the 1991 Pinatubo eruption as compared with the observed heating of only 1.0 K. The GCM calculations did not include consideration of the *quasi-biennial oscillation* (*QBO*) or O_3 effects. They then conducted another calculation to see the effects of the observed Pinatubo-induced O_3 depletion on stratospheric heating and found that it would have cut the calculated heating by 1 K. If the GCM had considered the observed QBO cooling and this O_3 effect, the calculated heating would agree quite well with observations.

The volcanic effect on O_3 chemistry is a new phenomenon, dependent on anthropogenic chlorine in the stratosphere. While we have no observations, the 1963 Agung eruption probably did not deplete O_3, as there was little anthropogenic chlorine in the stratosphere. Because of the Montreal protocol and subsequent international agreements, chlorine concentration has peaked in the stratosphere and is now decreasing. Therefore, for the next few decades, large volcanic eruptions will have effects similar to Pinatubo, but after that, these O_3 effects will go away and volcanic eruptions will have a stronger effect on atmospheric circulation without the negative feedback produced by O_3 depletion.

8. SUMMARY AND DISCUSSION

Large volcanic eruptions inject sulfur gases into the stratosphere, which convert to sulfate aerosols with an *e*-folding residence timescale of about 1 year. The climate response to large eruptions lasts for several years. The aerosol cloud produces cooling at the surface but heating in the stratosphere. For a tropical eruption this heating is larger in the tropics than in the high latitudes, producing an enhanced pole-to-equator temperature gradient and, in the Northern Hemisphere winter, a stronger polar vortex and winter warming of Northern Hemisphere continents. This indirect advective effect on temperature is stronger than the radiative cooling effect that dominates at lower latitudes and in the summer. The volcanic aerosols serve as surfaces for heterogeneous chemical reactions that destroy stratospheric ozone, which lowers ultraviolet absorption and reduces the radiative heating in the lower stratosphere. Since this chemical effect depends on the presence of anthropogenic chlorine, it has only become important in recent decades. There is no evidence that volcanic eruptions produce El Niño events, but the climatic consequences of El Niño and volcanic eruptions are similar and must be separated to understand the climatic response to each.

Because volcanic eruptions and their subsequent climatic response represent a large perturbation to the climate system over a relatively short period, observations and the simulated model responses can serve as important analogs for understanding the climatic response to other perturbations. While the climatic response to explosive volcanic eruptions is a useful analog for some other climatic forcings, there are also limitations.

The theory of "nuclear winter," the climatic effects of a massive injection of soot aerosols into the atmosphere from fires following a global nuclear holocaust [*Turco et al.*, 1983, 1990; *Robock*, 1984c, 1996b], includes upward injection of the aerosols to the stratosphere, rapid global dispersal of stratospheric aerosols, heating of the stratosphere, and cooling at the surface under this cloud. Because this theory cannot be tested in the real world, volcanic eruptions provide analogs that support these aspects of the theory.

Even though a climate model successfully simulates the response to a volcanic eruption [e.g., *Hansen et al.*, 1992], this does not guarantee that it can accurately simulate the response to greenhouse gases. The ability of the same model to respond to decadal- and longer-scale responses to climate forcings is not tested, as the interannual time-dependent response of the climate system depends only on the thermal inertia of the oceanic mixed layer combined with the climate model sensitivity. The long-term response to global warming depends on accurate simulation of the deep ocean circulation, and a volcano experiment does not allow an evaluation of that portion of the model. If a volcano experiment [e.g., *Kirchner et al.*, 1999] simulates the winter warming response of a climate model to volcanic aerosols, then we can infer the ability of a model to simulate the same dynamical mechanisms in response to increased greenhouse gases, ozone depletion, or solar variations.

As climate models improve, through programs like the GCM–Reality Intercomparison Project for the Stratosphere (GRIPS) of the Stratospheric Processes and their Relation to Climate (SPARC) program [*Pawson et al.*, 2000], they will improve their ability to simulate the climatic response to volcanic eruptions and other causes of climate change.

GLOSSARY

500-mbar pattern: Wind circulation in the middle of the troposphere, typically at a height of about 5.5 km, which is indicative of the general direction of transport of energy and moisture. Average surface pressure is about 1000 mbar, equal to 10^5 Pa.

Advective effect: Changes produced by horizontal transport of energy by winds as opposed to those caused by radiation.

Aerosol: A liquid droplet or solid particle suspended in a gas.

Apparent transmission: Amount of total solar radiation transmitted to the surface corrected for geometry and time of day of measurements.

December-January-February (DJF): Winter in the Northern Hemisphere.

e-folding decay time: The amount of time it takes for an anomaly to decay to e^{-1} of its initial perturbation. This is typically taken to be the definition of the time-scale of a logarithmic process.

Glass inclusions: Pockets of liquid or gas inside glass formed when molten rock cools after a volcanic eruption.

Heterogeneous chemical reactions: Chemical reactions that occur with interactions involving more than one phase. Gas reactions that occur on the surface of aerosol particles are an example.

Interdecadal: Varying from one decade (10-year period) to another.

Lithic: Pertaining to stone. In the context here, it means that the aerosol particles are solid and consist of crustal material.

Little Ice Age: The period 1500–1900 A.D., which was cooler than the period before or after.

Magmatic material: From the magma, the hot liquid from the interior of the Earth that often emerges during a volcanic eruption.

Medieval Warming: Relatively warm period, 900–1200 A.D.

Mt: Megaton = 10^6 T = 10^9 kg = 10^{12} g = teragram.

Optical depth: A measure of the amount of extinction of radiation along a path through the atmosphere, proportional to the amount of material in the path.

Pyrheliometer: Radiation measuring instrument that measures the direct solar flux.

Pyroclastic flow: A hot, ash-laden eruption cloud from a volcano that descends down the mountainside, in spite of its temperature, due to the weight of the suspended aerosols.

Quasi-biennial oscillation (QBO): A reversal of the wind direction in the tropical stratosphere that occurs approximately every 27 months and influences temperature and chemistry.

Stratospheric Aerosol and Gas Experiment II: A satellite instrument in space since 1984 that measures the amount of radiation in four visible and near-IR frequencies that is transmitted horizontally through different altitudes by limb scanning. The amount and size distribution of volcanic aerosols can be derived from these data.

Sulfate: Containing oxidized sulfur, in the form of SO_4. Sulfate aerosols consist of H_2O and H_2SO_4.

Teleconnection pattern: Pattern that is correlated but not contiguous.

Teragram: Equal to 10^{12} g = 10^9 kg = 10^6 T = megaton.

Total Ozone Mapping Spectrometer (TOMS): An instrument that has flown on satellites since 1979 and measures backscattered ultraviolet radiation in several wavelengths, from which the total column amount of O_3 and SO_2 can be derived.

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REFERENCES

Ahn, M. H., A new SO_2 retrieval algorithm using total ozone mapping spectrometer radiance and its applications, Ph.D. dissertation, 192 pp., Dep. of Meteorol., Univ. of Md., College Park, 1997.

Andronova, N. G., E. V. Rozanov, F. Yang, M. E. Schlesinger, and G. L. Stenchikov, Radiative forcing by volcanic aerosols from 1850 to 1994, *J. Geophys. Res.*, *104*, 16,807–16,826, 1999.

Angell, J. K., Impact of El Niño on the delineation of tropospheric cooling due to volcanic eruptions, *J. Geophys. Res.*, *93*, 3697–3704, 1988.

Angell, J. K., Estimated impact of Agung, El Chichón, and Pinatubo volcanic eruptions on global and regional total ozone after adjustment for the QBO, *Geophys. Res. Lett.*, *24*, 647–650, 1997a.

Angell, J. K., Stratospheric warming due to Agung, El Chichón, and Pinatubo taking into account the quasi-biennial oscillation, *J. Geophys. Res.*, *102*, 9479–9485, 1997b.

Asaturov, M. L., M. I. Budyko, K. Y. Vinnikov, P. Y. Groisman, A. S. Kabanov, I. L. Karol, M. P. Kolomeev, Z. I. Pivovarova, E. V. Rozanov, and S. S. Khmelevtsov, *Volcanics, Stratospheric Aerosol and Earth's Climate* (in Russian), 256 pp., Gidrometeoizdat, St. Petersburg, Russia, 1986.

Barkstrom, B. R., The Earth Radiation Budget Experiment (ERBE), *Bull. Am. Meteorol. Soc.*, *65*, 1170–1185, 1984.

Barnes, J. E., and D. J. Hoffman, Lidar measurements of stratospheric aerosol over Mauna Loa Observatory, *Geophys. Res. Lett.*, *24*, 1923–1926, 1997.

Bekki, S., J. A. Pyle, and D. M. Pyle, The role of microphysical and chemical processes in prolonging the climate forcing of the Toba eruption, *Geophys. Res. Lett.*, *23*, 2669–2672, 1996.

Bluth, G. J. S., S. D. Doiron, S. C. Schnetzler, A. J. Krueger, and L. S. Walter, Global tracking of the SO_2 clouds from the June 1991 Mount Pinatubo eruptions, *Geophys. Res. Lett.*, *19*, 151–154, 1992.

Bradley, R. S., The explosive volcanic eruption signal in Northern Hemisphere continental temperature records, *Clim. Change*, *12*, 221–243, 1988.

Bryson, R. A., and G. J. Dittberner, A non-equilibrium model of the hemisphere, *J. Atmos. Sci.*, *33*, 2094–2106, 1976.

Coffey, M. T., Observations of the impact of volcanic activity on stratospheric chemistry, *J. Geophys. Res.*, *101*, 6767–6780, 1996.

Crowley, T. J., and K.-Y. Kim, Modeling the temperature response to forced climate change over the last six centuries, *Geophys. Res. Lett.*, *26*, 1901–1904, 1999.

Crutzen, P. J., The possible importance of CSO for the sulfate layer of the stratosphere, *Geophys. Res. Lett.*, *3*, 73–76, 1976.

D'Arrigo, R. D., G. C. Jacoby, M. Free, and A. Robock, Northern Hemisphere temperature variability for the past three centuries: Tree-ring and model estimates, *Clim. Change*, *42*, 663–675, 1999.

Dutton, E. G., A coherence between the QBO and the amplitude of the Mauna Loa atmospheric transmission annual cycle, *Int. J. Climatol.*, *12*, 383–396, 1992.

Dutton, E. G., and J. R. Christy, Solar radiative forcing at selected locations and evidence for global lower tropospheric cooling following the eruption of El Chichón and Pinatubo, *Geophys. Res. Lett.*, *19*, 2313–2316, 1992.

Dutton, E. G., J. J. DeLuisi, and A. P. Austring, Interpretation of Mauna Loa atmospheric transmission relative to aerosols, using photometric precipitable water amounts, *J. Atmos. Chem.*, *3*, 53–68, 1985.

Ellsaesser, H. W., Isolating the climatologic effects of volcanoes, *Rep. UCRL-89161*, 29 pp., Lawrence Livermore Natl. Lab., Livermore, Calif., 1983.

Forsyth, P. Y., In the wake of Etna, 44 B.C., *Classical Antiq.*, *7*, 49–57, 1988.

Franklin, B., Meteorological imaginations and conjectures, *Manchr. Lit. Philos. Soc. Mem. Proc.*, *2*, 122, 1784. (Reprinted in *Weatherwise*, *35*, 262, 1982.)

Free, M., and A. Robock, Global warming in the context of the Little Ice Age, *J. Geophys. Res.*, *104*, 19,057–19,070, 1999.

Gates, W. L., AMIP: The Atmospheric Model Intercomparison Project, *Bull. Am. Meteorol. Soc.*, *73*, 1962–1970, 1992.

Genin, A., B. Lazar, and S. Brenner, Vertical mixing and coral death in the Red Sea following the eruption of Mount Pinatubo, *Nature*, *377*, 507–510, 1995.

Gilliland, R. L., Solar, volcanic, and CO_2 forcing of recent climatic changes, *Clim. Change*, *4*, 111–131, 1982.

Gilliland, R. L., and S. H. Schneider, Volcanic, CO_2, and solar forcing of Northern and Southern Hemisphere surface temperatures, *Nature*, *310*, 38–41, 1984.

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

Graf, H.-F., J. Perlwitz, I. Kirchner, and I. Schult, On the interrelationship between recent climate trends, ozone change and increased greenhouse gas forcing, in *Atmospheric Ozone As a Climate Gas*, edited by W. C. Wang and I. S. A. Isaksen, pp. 163–179, Springer-Verlag, New York, 1995.

Graf, H.-F., J. Feichter, and B. Langmann, Volcanic sulfur emission: Estimates of source strength and its contribution to the global sulfate distribution, *J. Geophys. Res.*, *102*, 10,727–10,738, 1997.

Grant, W. B., Tropical stratospheric ozone changes following the eruption of Mount Pinatubo, in *The Mount Pinatubo Eruption Effects on the Atmosphere and Climate*, *NATO ASI Ser.*, *Ser. I*, *42*, 161–175, 1996.

Grant, W. B., et al., Observations of reduced ozone concentrations in the tropical atmosphere after the eruption of Mount Pinatubo, *Geophys. Res. Lett.*, *19*, 1109–1112, 1992.

Grattan, J., M. Brayshay, and J. Sadler, Modelling the distal impacts of past volcanic gas emissions: Evidence of Europe-wide environmental impacts from gases emitted during the eruption of Italian and Icelandic volcanoes in 1783, *Quaternaire*, *9*, 25–35, 1998.

Groisman, P. Y., Regional climate consequences of volcanic eruptions (in Russian), *Meteorol. Hydrol.*, No. 4, 39–45, 1985.

Groisman, P. Y., Possible regional climate consequences of the Pinatubo eruption: An empirical approach, *Geophys. Res. Lett.*, *19*, 1603–1606, 1992.

Haigh, J., The impact of solar variability on climate, *Science*, *272*, 981–984, 1996.

Halpert, M. S., and C. F. Ropelewski, Surface temperature patterns associated with the Southern Oscillation, *J. Clim.*, *5*, 577–593, 1992.

Handler, P., Possible association between the climatic effects of stratospheric aerosols and sea surface temperatures in the eastern tropical Pacific Ocean, *J. Climatol.*, *6*, 31–41, 1986.

Hansen, J. E., W.-C. Wang, and A. A. Lacis, Mount Agung provides a test of a global climatic perturbation, *Science*, *199*, 1065–1068, 1978.

Hansen, J., I. Fung, A. Lacis, D. Rind, S. Lebedeff, R. Ruedy, G. Russell, and P. Stone, Global climate changes as forecast by the Goddard Institute for Space Studies three-dimensional model, *J. Geophys. Res.*, *93*, 9341–9364, 1988.

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., et al., A Pinatubo climate modeling investigation, in *The Mount Pinatubo Eruption Effects on the Atmosphere and Climate*, *NATO ASI Ser.*, *Ser. I*, *42*, 233–272, 1996.

Harington, C. R. (Ed.), *The Year Without a Summer? World Climate in 1816*, 576 pp., Can. Mus. of Nature, Ottawa, Ont., Canada, 1992.

Harshvardhan, Perturbations of the zonal radiation balance by a stratospheric aerosol layer, *J. Atmos. Sci.*, *36*, 1274–1285, 1979.

Hirono, M., On the trigger of El Niño–Southern Oscillation by the forcing of early El Chichón volcanic aerosols, *J. Geophys. Res.*, *93*, 5365–5384, 1988.

Holton, J. R., P. H. Haynes, M. E. McIntyre, A. R. Douglass, R. B. Rood, and L. Pfister, Stratosphere-troposphere exchange, *Rev. Geophys.*, *33*, 403–439, 1995.

Houghton, J. T., L. G. Meira Filho, B. A. Callander, N. Harris, A. Kattenberg, and K. Maskell (Eds.), *Climate Change 1995—The Science of Climate Change*, 584 pp., Cambridge Univ. Press, New York, 1996.

Humphreys, W. J., Volcanic dust and other factors in the production of climatic changes, and their possible relation to ice gases, *J. Franklin Inst.*, Aug., 131–172, 1913.

Humphreys, W. J., *Physics of the Air*, 676 pp., Dover, Mineola, N. Y., 1940.

Hunt, B. G., A simulation of the possible consequences of a volcanic eruption on the general circulation of the atmosphere, *Mon. Weather Rev.*, *105*, 247–260, 1977.

Hurrell, J. W., Decadal trends in the North Atlantic Oscillation: Regional temperatures and precipitation, *Science*, *269*, 676–679, 1995.

Hurrell, J. W., and H. van Loon, Decadal variations in climate associated with the North Atlantic Oscillation, *Clim. Change*, *36*, 301–326, 1997.

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

Keen, R. A., Volcanic aerosols and lunar eclipses, *Science*, *222*, 1011–1013, 1983.

Kelly, P. M., and C. B. Sear, The formulation of Lamb's dust veil index, in *Atmospheric Effects and Potential Climatic Impact of the 1980 Eruptions of Mount St. Helens*, edited by A. Deepak, *NASA Conf. Publ. 2240*, 293–298, 1982.

Kelly, P. M., P. D. Jones, A. Robock, and K. R. Briffa, The contribution of Hubert H. Lamb to the study of volcanic effects on climate, *Weather*, *53*, 209–222, 1998.

Kinne, S., O. B. Toon, and M. J. Prather, Buffering of stratospheric circulation by changing amounts of tropical ozone: A Pinatubo case study, *Geophys. Res. Lett.*, *19*, 1927–1930, 1992.

Kirchner, I., and H.-F. Graf, Volcanos and El Niño: Signal separation in Northern Hemisphere winter, *Clim. Dyn.*, *11*, 341–358, 1995.

Kirchner, I., G. L. Stenchikov, H.-F. Graf, A. Robock, and J. C. Antuña, Climate model simulation of winter warming and summer cooling following the 1991 Mount Pinatubo volcanic eruption, *J. Geophys. Res.*, *104*, 19,039–19,055, 1999.

Kitoh, A., H. Koide, K. Kodera, S. Yukimoto, and A. Noda, Interannual variability in the stratospheric-tropospheric circulation in a coupled ocean-atmosphere GCM, *Geophys. Res. Lett.*, *23*, 543–546, 1996.

Kodera, K., Influence of volcanic eruptions on the troposphere through stratospheric dynamical processes in the Northern Hemisphere winter, *J. Geophys. Res.*, *99*, 1273–1282, 1994.

Kodera, K., M. Chiba, and K. Shibata, A general circulation model study of the solar and QBO modulation of the stratospheric circulation during the Northern Hemisphere winter, *Geophys. Res. Lett.*, *18*, 1209–1212, 1991.

Kodera, K., M. Chiba, H. Koide, A. Kitoh, and Y. Nikaidou, Interannual variability of the winter stratosphere and troposphere in the Northern Hemisphere, *J. Meteorol. Soc. Jpn.*, *74*, 365–382, 1996.

Kondratyev, K. Y., Volcanoes and climate, *WCP-54*, *WMO/TD-166*, 103 pp., World Meteorol. Org., Geneva, 1988.

Kondratyev, K. Y., and I. Galindo, *Volcanic Activity and Climate*, 382 pp., A. Deepak, Hampton, Va., 1997.

Krueger, A. J., C. C. Schnetzler, and L. S. Walter, The December 1981 eruption of Nyamuragira Volcano (Zaire) and the origin of the "mystery cloud" of early 1982, *J. Geophys. Res.*, *101*, 15,191–15,196, 1996.

Labitzke, K., and H. van Loon, Association between the 11-year solar cycle, the QBO, and the atmosphere, part I, The troposphere and the stratosphere in the Northern Hemisphere in winter, *J. Atmos. Terr. Phys.*, *50*, 197–207, 1988.

Lamb, H. H., Volcanic dust in the atmosphere, with a chronology and assessment of its meteorological significance, *Philos. Trans. R. Soc. London*, *Ser. A*, *266*, 425–533, 1970.

Lamb, H. H., Supplementary volcanic dust veil index assessments, *Clim. Monit.*, *6*, 57–67, 1977.

Lamb, H. H., Update of the chronology of assessments of the volcanic dust veil index, *Clim. Monit.*, *12*, 79–90, 1983.

Li, X. S., and A. Berger, Simulating climatic impact of the Toba eruption with the LLN 2-D climate model, *Sci. Rep. 1997/14*, Inst. d'Astron. et de Geophys. G. Lemaitre, Univ. Cath. de Louvain, Louvain-la-Neuve, Belgium, 1997.

Lough, J. M., and H. C. Fritts, An assessment of the possible effects of volcanic eruptions on North American climate using tree-ring data, 1602 to 1900 A.D., *Clim. Change*, *10*, 219–239, 1987.

Luhr, J. F., I. S. E. Carmichael, and J. C. Varekamp, The 1982 eruptions of El Chichón volcano, Chiapas, Mexico: Mineralogy and petrology of the anhydrite-bearing pumices, *J. Volcanol. Geotherm. Res.*, *23*, 69–108, 1984.

MacCracken, M. C., and F. M. Luther, Preliminary estimate of the radiative and climatic effects of the El Chichón eruption, *Geofis. Int.*, *23*, 385–401, 1984.

Mao, J., and A. Robock, Surface air temperature simulations by AMIP general circulation models: Volcanic and ENSO signals and systematic errors, *J. Clim.*, *11*, 1538–1552, 1998.

Marcus, S. L., Y. Chao, J. O. Dickey, and P. Gegout, Detection and modeling of nontidal oceanic effects on Earth's rotation rate, *Science*, *281*, 1656–1659, 1998.

Mass, C. F., and D. A. Portman, Major volcanic eruptions and climate: A critical evaluation, *J. Clim.*, *2*, 566–593, 1989.

Mass, C., and A. Robock, The short-term influence of the Mount St. Helens volcanic eruption on surface temperature in the northwest United States, *Mon. Weather Rev.*, *110*, 614–622, 1982.

McIntyre, M. E., How well do we understand the dynamics of stratospheric warmings?, *J. Meteorol. Soc. Jpn.*, *60*, 37–65, 1982.

Miles, M. K., and P. B. Gildersleeves, Volcanic dust and changes in Northern Hemisphere temperature, *Nature*, *271*, 735–736, 1978.

Minnis, P., E. F. Harrison, L. L. Stowe, G. G. Gison, F. M. Denn, D. R. Doelling, and W. L. Smith Jr., Radiative climate forcing by the Mount Pinatubo eruption, *Science*, *259*, 1411–1415, 1993.

Mitchell, J. M., Jr., Recent secular changes of the global temperature, *Ann. N. Y. Acad. Sci.*, *95*, 235–250, 1961.

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.

Mohnen, V. A., Stratospheric ion and aerosol chemistry and possible links with cirrus cloud microphysics—A critical assessment, *J. Atmos. Sci.*, *47*, 1933–1948, 1990.

Newhall, C. G., and S. Self, The volcanic explosivity index (VEI): An estimate of explosive magnitude for historical volcanism, *J. Geophys. Res.*, *87*, 1231–1238, 1982.

Nicholls, N., Low latitude volcanic eruptions and the El Niño–Southern Oscillation, *J. Climatol.*, *8*, 91–95, 1988.

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

Parker, D. E., and J. K. L. Brownscombe, Stratospheric warming following the El Chichón volcanic eruption, *Nature*, *301*, 406–408, 1983.

Pawson, S., et al., The GCM-Reality Intercomparison Project for SPARC (GRIPS): Scientific issues and initial results, *Bull. Am. Meteorol. Soc.*, *81*, 781–796, 2000.

Perlwitz, J., and H.-F. Graf, The statistical connection between tropospheric and stratospheric circulation of the Northern Hemisphere in winter, *J. Clim.*, *8*, 2281–2295, 1995.

Pinto, J. P., R. P. Turco, and O. B. Toon, Self-limiting physical and chemical effects in volcanic eruption clouds, *J. Geophys. Res.*, *94*, 11,165–11,174, 1989.

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.

Pollack, J. B., D. Rind, A. Lacis, J. E. Hansen, M. Sato, and R. Ruedy, GCM simulation of volcanic aerosol forcing, part I, Climate changes induced by steady-state perturbations, *J. Clim.*, *6*, 1719–1742, 1993.

Quiroz, R. S., The isolation of stratospheric temperature change due to the El Chichón volcanic eruption from nonvolcanic signals, *J. Geophys. Res.*, *88*, 6773–6780, 1983.

Ramaswamy, V., M. D. Schwarzkopf, and W. J. Randel, Fingerprint of ozone depletion in the spatial and temporal pattern of recent lower-stratospheric cooling, *Nature*, *382*, 616–618, 1996.

Rampino, M. R., and S. Self, Sulphur-rich volcanic eruptions and stratospheric aerosols, *Nature*, *310*, 677–679, 1984.

Rampino, M. R., and S. Self, Volcanic winter and accelerated glaciation following the Toba super-eruption, *Nature*, *359*, 50–52, 1992.

Rampino, M. R., S. Self, and R. W. Fairbridge, Can climatic change cause volcanic eruptions?, *Science*, *206*, 826–828, 1979.

Rind, D., N. K. Balachandran, and R. Suozzo, Climate change

and the middle atmosphere, part II, The impact of volcanic aerosols, *J. Clim.*, *5*, 189–208, 1992.
Robock, A., Internally and externally caused climate change, *J. Atmos. Sci.*, *35*, 1111–1122, 1978.
Robock, A., The "Little Ice Age": Northern Hemisphere average observations and model calculations, *Science*, *206*, 1402–1404, 1979.
Robock, A., The Mount St. Helens volcanic eruption of 18 May 1980: Minimal climatic effect, *Science*, *212*, 1383–1384, 1981a.
Robock, A., A latitudinally dependent volcanic dust veil index, and its effect on climate simulations, *J. Volcanol. Geotherm. Res.*, *11*, 67–80, 1981b.
Robock, A., The dust cloud of the century, *Nature*, *301*, 373–374, 1983a.
Robock, A., Ice and snow feedbacks and the latitudinal and seasonal distribution of climate sensitivity, *J. Atmos. Sci.*, *40*, 986–997, 1983b.
Robock, A., Review of *Volcano Weather: The Story of 1816, the Year Without a Summer*, *Natl. Weather Dig.*, *9*, 5–6, 1984a.
Robock, A., Climate model simulations of the effects of the El Chichón eruption, *Geofis. Int.*, *23*, 403–414, 1984b.
Robock, A., Snow and ice feedbacks prolong effects of nuclear winter, *Nature*, *310*, 667–670, 1984c.
Robock, A., Volcanoes and climate, in *Climate and Geo-Sciences: A Challenge for Science and Society in the 21st Century*, *NATO ASI Ser., Ser. C*, *285*, 309–314, 1989.
Robock, A., The volcanic contribution to climate change of the past 100 years, in *Greenhouse-Gas-Induced Climatic Change: A Critical Appraisal of Simulations and Observations*, edited by M. E. Schlesinger, pp. 429–444, Elsevier Sci., New York, 1991.
Robock, A., Review of *Year Without a Summer? World Climate in 1816*, *Clim. Change*, *26*, 105–108, 1994.
Robock, A., Stratospheric control of climate, *Science*, *272*, 972–973, 1996a.
Robock, A., Nuclear winter, in *Encyclopedia of Weather and Climate*, vol. 2, edited by S. H. Schneider, pp. 534–536, Oxford Univ. Press, New York, 1996b.
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 M. P. Free, The volcanic record in ice cores for the past 2000 years, in *Climatic Variations and Forcing Mechanisms of the Last 2000 Years*, edited by P. D. Jones, R. S. Bradley, and J. Jouzel, pp. 533–546, Springer-Verlag, New York, 1996.
Robock, A., and Y. Liu, The volcanic signal in Goddard Institute for Space Studies three-dimensional model simulations, *J. Clim.*, *7*, 44–55, 1994.
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.
Robock, A., and C. Mass, The Mount St. Helens volcanic eruption of 18 May 1980: Large short-term surface temperature effects, *Science*, *216*, 628–630, 1982.
Robock, A., and M. Matson, Circumglobal transport of the El Chichón volcanic dust cloud, *Science*, *221*, 195–197, 1983.
Robock, A., K. E. Taylor, G. L. Stenchikov, and Y. Liu, GCM evaluation of a mechanism for El Niño triggering by the El Chichón ash cloud, *Geophys. Res. Lett.*, *22*, 2369–2372, 1995.
Roeckner, E., K. Arpe, L. Bengtsson, M. Christoph, M. Claussen, L. Duemenil, M. Ecsh, M. Giogetta, U. Schlese, and U. Schulzweida, The atmospheric general circulation model ECHAM4: Model description and simulation of present-day climate, *Rep. 218*, 90 pp., Max-Planck-Inst. für Meteorol., Hamburg, Germany, 1996.
Ropelewski, C. F., and P. D. Jones, An extension of the Tahiti-Darwin Southern Oscillation Index, *Mon. Weather Rev.*, *115*, 2161–2165, 1987.
Rosenfield, J., D. Considine, P. Meade, J. Bacmeister, C. Jackman, and M. Schoeberl, Stratospheric effects of Mount Pinatubo aerosol studied with a coupled two-dimensional model, *J. Geophys. Res.*, *102*, 3649–3670, 1997.
Russell, P. B., et al., Global to microscale evolution of the Pinatubo volcanic aerosol derived from diverse measurements and analyses, *J. Geophys. Res.*, *101*, 18,745–18,763, 1996.
Sapper, K., Beiträge zur Geographie der tätingen Vulkane, *Z. Vulkanol.*, *3*, 65–197, 1917.
Sapper, K., *Vulkankunde*, Englehorn Verlag, Stuttgart, Germany, 1927.
Sassen, K., D. O. Starr, G. G. Mace, M. R. Poellot, S. H. Melfi, W. L. Eberhard, J. D. Spinhirne, E. W. Eloranta, D. E. Hagen, and J. Hallett, The 5–6 December 1991 FIRE IFO II jet stream cirrus case study: Possible influences of volcanic aerosols, *J. Atmos. Sci.*, *52*, 97–123, 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.
Schatten, K. H., H. G. Mayr, I. Harris, and H. A. Taylor, A zonally symmetric model for volcanic influence upon atmospheric circulation, *Geophys. Res. Lett.*, *11*, 303–306, 1984.
Schneider, S. H., and C. Mass, Volcanic dust, sunspots and temperature trends, *Science*, *190*, 741–746, 1975.
Self, S., M. R. Rampino, J. Zhao, and M. G. Katz, Volcanic aerosol perturbations and strong El Niño events: No general correlation, *Geophys. Res. Lett.*, *24*, 1247–1250, 1997.
Shelley, M., *Frankenstein*, Bantam Books, New York, 1818. (Classic edition, 209 pp., Bantam Books, New York, 1981.)
Shindell, D., D. Rind, N. Balachandran, J. Lean, and P. Lonergan, Solar cycle variability, ozone, and climate, *Science*, *284*, 305–308, 1999.
Simarski, L. T., Volcanism and climate change, special report, 27 pp., AGU, Washington, D. C., 1992.
Simkin, T., and R. S. Fiske, *Krakatau 1883: The Volcanic Eruption and Its Effects*, 464 pp., Smithson. Inst., Washington, D. C., 1983.
Simkin, T., and L. Siebert, *Volcanoes of the World*, 2nd ed., 349 pp., Geoscience Press, Tucson, Ariz., 1994.
Simkin, T., L. Siebert, L. McClelland, D. Bridge, C. G. Newhall, and J. H. Latter, *Volcanoes of the World*, 232 pp., Van Nostrand Reinhold, New York, 1981.
Solomon, S., Stratospheric ozone depletion: A review of concepts and history, *Rev. Geophys.*, *37*, 275–316, 1999.
Solomon, S., R. W. Portmann, R. R. Garcia, L. W. Thomason, L. R. Poole, and M. P. McCormick, The role of aerosol variations in anthropogenic ozone depletion at northern midlatitudes, *J. Geophys. Res.*, *101*, 6713–6727, 1996.
Spencer, R. W., J. R. Christy, and N. C. Grody, Global atmospheric temperature monitoring with satellite microwave measurements: Method and results 1979–1984, *J. Clim.*, *3*, 1111–1128, 1990.
Stenchikov, G. L., I. Kirchner, A. Robock, H.-F. Graf, J. C. Antuña, R. G. Grainger, A. Lambert, and L. Thomason, Radiative forcing from the 1991 Mount Pinatubo volcanic eruption, *J. Geophys. Res.*, *103*, 13,837–13,857, 1998.
Stommel, H., and E. Stommel, *Volcano Weather: The Story of 1816, the Year Without a Summer*, 177 pp., Seven Seas Press, Newport, R. I., 1983.
Stothers, R. B., The great Tambora eruption in 1815 and its aftermath, *Science*, *224*, 1191–1198, 1984.
Stowe, L. L., R. M. Carey, and P. P. Pellegrino, Monitoring the Mount Pinatubo aerosol layer with NOAA 11 AVHRR data, *Geophys. Res. Lett.*, *19*, 159–162, 1992.
Strong, A. E., Monitoring El Chichón aerosol distribution using NOAA-7 satellite AVHRR sea surface temperature observations, *Geofis. Int.*, *23*, 129–141, 1984.

Strong, A. E., The effect of El Chichón on the 82/83 El Niño (abstract), *Eos Trans. AGU*, *67*(44), 880, 1986.

Symons, G. J. (Ed.), *The Eruption of Krakatoa, and Subsequent Phenomena*, 494 pp., Trübner, London, 1888.

Tabazadeh, A., and R. P. Turco, Stratospheric chlorine injection by volcanic eruptions: HCl scavenging and implication for ozone, *Science*, *260*, 1082–1086, 1993.

Thompson, D. W. J., and J. M. Wallace, The Arctic Oscillation signature in the wintertime geopotential height and temperature fields, *Geophys. Res. Lett.*, *25*, 1297–1300, 1998.

Thompson, D. W. J., and J. M. Wallace, Annular modes in the extratropical circulation, part I, Month-to-month variability, *J. Clim.*, *13*, 1000–1016, 2000a.

Thompson, D. W. J., J. M. Wallace, and G. C. Hegerl, Annular modes in the extratropical circulation, part II, Trends, *J. Clim.*, *13*, 1018–1036, 2000b.

Tie, X., and G. Brasseur, The response of the stratospheric ozone to volcanic eruptions: Sensitivity to atmospheric chlorine loading, *Geophys. Res. Lett.*, *22*, 3035–3038, 1995.

Tie, X., G. P. Brasseur, C. Granier, A. De Rudder, and N. Larsen, Model study of polar stratospheric clouds and their effect on stratospheric ozone, 2, Model results, *J. Geophys. Res.*, *101*, 12,575–12,584, 1996.

Toon, O. B., Volcanoes and climate, in *Atmospheric Effects and Potential Climatic Impact of the 1980 Eruptions of Mount St. Helens*, edited by A. Deepak, *NASA Conf. Publ. 2240*, 15–36, 1982.

Toon, O. B., and J. B. Pollack, Atmospheric aerosols and climate, *Am. Sci.*, *68*, 268–278, 1980.

Trepte, C. R., and M. H. Hitchman, Tropical stratospheric circulation deduced from satellite aerosol data, *Nature*, *355*, 626–628, 1992.

Trepte, C. R., R. E. Veiga, and M. P. McCormick, The poleward dispersal of Mount Pinatubo volcanic aerosol, *J. Geophys. Res.*, *98*, 18,563–18,573, 1993.

Turco, R. P., O. B. Toon, T. P. Ackerman, J. B. Pollack, and C. Sagan, Nuclear winter: Global consequences of multiple nuclear explosions, *Science*, *222*, 1283–1292, 1983.

Turco, R. P., O. B. Toon, T. P. Ackerman, J. B. Pollack, and C. Sagan, Nuclear winter: Climate and smoke: An appraisal of nuclear winter, *Science*, *247*, 166–176, 1990.

van Loon, H., and K. Labitzke, Association between the 11-year solar cycle, the QBO and the atmosphere, part IV, The stratosphere, not grouped by the phase of the QBO, *J. Clim.*, *3*, 827–837, 1990.

Vinnikov, K. Y., A. Robock, R. J. Stouffer, and S. Manabe, Vertical patterns of free and forced climate variations, *Geophys. Res. Lett.*, *23*, 1801–1804, 1996.

Vogelmann, A. M., T. P. Ackerman, and R. P. Turco, Enhancements in biologically effective ultraviolet radiation following volcanic eruptions, *Nature*, *359*, 47–49, 1992.

Vupputuri, R. K. R., and J. P. Blanchet, The possible effects of the El Chichón eruption on atmospheric thermal and chemical structure and surface climate, *Geofis. Int.*, *23*, 433–447, 1984.

Wallace, J. M., and D. S. Gutzler, Teleconnections in the geopotential height field during the Northern Hemisphere winter, *Mon. Weather Rev.*, *109*, 784–812, 1981.

Wallace, J. M., Y. Zhang, and J. A. Renwick, Dynamic contribution to hemispheric mean temperature trends, *Science*, *270*, 780–783, 1995.

Wallace, J. M., Y. Zhang, and L. Balik, Interpretation of interdecadal trends in Northern Hemisphere surface air temperature, *J. Clim.*, *9*, 249–259, 1996.

Yue, G. K., R. E. Veiga, and P.-H. Wang, SAGE II observations of a previously unreported stratospheric volcanic aerosol cloud in the northern polar summer of 1980, *Geophys. Res. Lett.*, *21*, 429–432, 1994.

Zerefos, C. S., K. Tourpali, and A. F. Bais, Further studies on possible volcanic signal to the ozone layer, *J. Geophys. Res.*, *99*, 25,741–25,746, 1994.

Zhao, J., R. P. Turco, and O. B. Toon, A model simulation of Pinatubo volcanic aerosols in the stratosphere, *J. Geophys. Res.*, *100*, 7315–7328, 1995.

Zielinski, G. A., P. A. Mayewski, L. D. Meeker, S. Whitlow, M. S. Twickler, and K. Taylor, Potential atmospheric impact of the Toba mega-eruption ~71,000 years ago, *Geophys. Res. Lett.*, *23*, 837–840, 1996.

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