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VOLCANIC WINTERS[1]

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INTRODUCTION: THE VOLCANO/CLIMATE CONNECTION

Accounts of prolonged darkness, often associated with abnormally cold weather and hardship, are common in the myths and legends of many cultures. Egyptian papyruses corroborate the statement in the Book of Exodus in the Bible that "there was a thick darkness in all the land of Egypt for three days." Similar kinds of stories can be found in ancient Sumerian, Greek, and Mayan literature. These have on occasion been used to argue for global catastrophes, such as encounters or collisions with comets, in early historical times. A more reasonable explanation comes from the similarity of these reports to more recent historical accounts of the aftereffects of large volcanic eruptions. Thus, a substantial case has been made for the connection of the Egyptian and Biblical reports of darkness and ash rains at the time of the Exodus with the explosive eruption of Santorini (Thera) in the Aegean Sea in the second millenium BC (see references in Downey & Tarling 1984, Stanley & Sheng 1986), which also somehow contributed to the demise of Minoan Crete (Marinatos 1939) and had effects as far away as China (Pang & Chou 1985). In more recent times, the inhabitants of interior New Guinea speak of the "time of darkness," a tradition passed down over generations since the seventeenth century AD. Blong (1982) has shown that this darkness was accompanied by large local temperature changes, both hot and cold, and was most likely related to the ash cloud from an eruption of Long Island, one of the active volcanoes off the northeast coast of New Guinea.

Considering the knowledge that we have gained in the past few decades about the mechanisms of volcanic eruptions and the generation of volcanic aerosol clouds in the atmosphere, it is reasonable to ask what the atmospheric effects of the largest eruptions might be. Beyond the local devastation and regional effects, it is known that some historical eruptions had a noticeable impact on climate and agriculture on a hemispheric to global basis. This being the case, much larger eruptions may possibly have caused severe "volcanic winters," perhaps similar to the recently proposed "nuclear winter." These "supereruptions" must therefore be considered in discussions of natural hazards that might have global consequences (Rampino et al 1985, Burke & Francis 1985, Smith 1985).

Modern interest in the problem of the impact of volcanic eruptions on the atmosphere and climate is traditionally traced back to the observations of Benjamin Franklin at the time of the eruption of Laki (Lakagígar) in Iceland in 1783. Franklin (see Lamb 1970, p. 433) described what he termed a "dry fog" in Europe during his stay there as minister to France. As Franklin wrote, the rays of the Sun "were indeed rendered so faint in passing through it that, when collected in the focus of a burning glass, they would scarce kindle brown paper." Franklin connected the dry fog and the reduced solar radiation with the severe winter of 1783/1784 in Europe and eastern North America, and he proposed that the Icelandic eruption of that time was to blame.

The connection between large volcanic eruptions and worldwide perturbations of the optical properties of the atmosphere was established by the classic study of the Krakatoa Commission (Russell & Archibald 1888) in the aftermath of the changes in the atmosphere seen after the Krakatau eruption in 1883 (see also Simkin & Fiske 1983). A number of atmospheric optical phenomena were identified, including noticeable dimming and blurring of celestial objects, unusual blue or green color of the Sun and Moon, enhanced sunrises and sunsets with lavender glows that appeared high over the horizon, Bishop's rings (a complex halo around the Sun produced by diffraction of sunlight by small particles, in which the normal order of colors is reversed, with red on the outside), and also unusually dark lunar eclipses (Flammarion 1884). A link between climate change and volcanic eruptions was also based on later studies that suggested a possible correlation between some eruptions, decreases in solar radiation measured at ground observatories, and short-term coolings of the Earth's surface (Humphreys 1913, Abbot & Fowle 1913; for an early review, see Humphreys 1940); however, in other similar studies the same or other noteworthy eruptions seemed to show no noticeable effect on the global climate (Gentilli 1948, Deirmendjian 1973, Landsberg & Albert 1974, Ellsaesser 1986).

In 1970, a classic study by H. H. Lamb clearly presented the empirical evidence for a volcano/climate connection as understood at that time. Lamb reviewed previous work on the subject, but his most valuable contribution was his tabulation of a chronology of important volcanic eruptions for the period subsequent to AD 1500 and his definition of the volcanic dust veil index (d.v.i.), an estimate of the amount of fine volcanic ash or dust lofted into the upper atmosphere by specific historical eruptions. Lamb concluded that some significant correlations existed between "volcanic eruption years" with high d.v.i. values and climatic cooling, but he stressed that in some cases where the d.v.i. was assessed largely on the evidence of temperature variation, one was in clear danger of arguing in a circle when investigating the possible effects of eruptions on climate. Newhall & Self (1982; see also Simkin et al 1981) attempted to further quantify volcanic "explosivity" in their Volcanic Explosivity Index (VEI), which combined estimates of eruption volume with explosive energy as evidenced by the height of the eruption plume. Hirschboeck (1980) proposed another, but simpler, index based largely on the volume of the eruption. By that time, however, it was clear that the composition of the volcanic ejecta, particularly the amount of sulfur volatiles released, had an importance above and beyond that of the total amount of ash ejected. The geographic location, time of year, and prevailing climatic conditions (e.g. phase in the quasi-biennial oscillation cycle) were also seen to be critical factors in determining the spread and lifetime of volcanic aerosol clouds.

NINETEENTH AND TWENTIETH CENTURY ERUPTIONS: SULFUR IS THE KEY

Although most workers prior to the late 1960s stressed the importance of "volcanic ash" in the stratospheric clouds (e.g. Jacobs 1954, Mitchell 1961), Lamb (1970) and Deirmendjian (1973) both recognized a possible connection between sulfur gases (primarily SO_2 and H_2S) injected into the upper atmosphere by volcanic eruptions and the evidence discovered by Junge et al (1961) for a supposedly permanent layer of sulfate aerosols (consisting largely of small droplets of sulfuric acid) in the stratosphere at around 25 km. It was later shown that the majority of the H_2SO_4 aerosols in the so-called Junge Layer were volcanic in origin (Castleman et al 1974).

A number of studies followed that focused on photochemical reactions and nucleation of sulfuric acid aerosols in the lower stratosphere. It was soon well established that the bulk of volcanic "dust" veils consisted of fine droplets of sulfuric acid [see Turco et al (1982) for a review]. Most of the volcanic ash fell out of the stratosphere in a few months, while the aerosols continued to nucleate and grow, creating the volcanic cloud that spread over wide areas of the globe and persisted for several years (Pollack et al 1976, Cadle et al 1976, 1977, Hunt 1977, Capone et al 1983). In addition, HCl and water vapor that are injected into the stratosphere during an eruption may have significant effects on the ozone concentrations (Hofmann 1987, Pinto et al 1987).

Stratospheric aerosols affect the global radiation budget by absorbing and, more importantly, backscattering incoming solar radiation (although they also absorb some outgoing infrared radiation from the ground). Absorption and backscattering of solar radiation should cause a cooling of the lower atmosphere and the surface. The absorption of infrared radiation should also cause an increase in the stratospheric temperatures (see Turco et al 1982). The volcanic signal expected in hemispheric or zonal surface temperature records in historical times, however, is about the same as the background interannual variations in temperature. Several studies have made use of the method of "compositing" or "superposed epoch analysis," in which the temperature records of several years bracketing a number of different eruptions are superposed in order to strengthen the contrast between the possible volcanic signal and background noise (Figure 1). These studies identified a statistically significant average temperature decrease of about 0.2 to 0.5°C for 1 to 3 years following the times of known nineteenth and twentieth century eruptions (Mitchell 1961, Mass & Schneider 1977, Taylor et al 1980, Self et al 1981, Angell & Korshover 1985, Lough & Fritts 1987). Other statistical studies have come to similar conclusions regarding the magnitude and duration of cooling after volcanic eruptions of the past 100 years (for a review, see Angell & Korshover 1985). A recent superposed epoch analysis of Northern Hemisphere sea-surface temperatures after major eruptions of the last 100 years, however, did not show any consistent response of posteruption cooling (Parker 1985), but sea-surface temperatures are expected to be less responsive to short-term temperature perturbations, and other factors such as the Southern Oscillation/El Niño phenomenon may be masking the volcanic climate signal.

Another approach is to focus on the largest and/or best documented

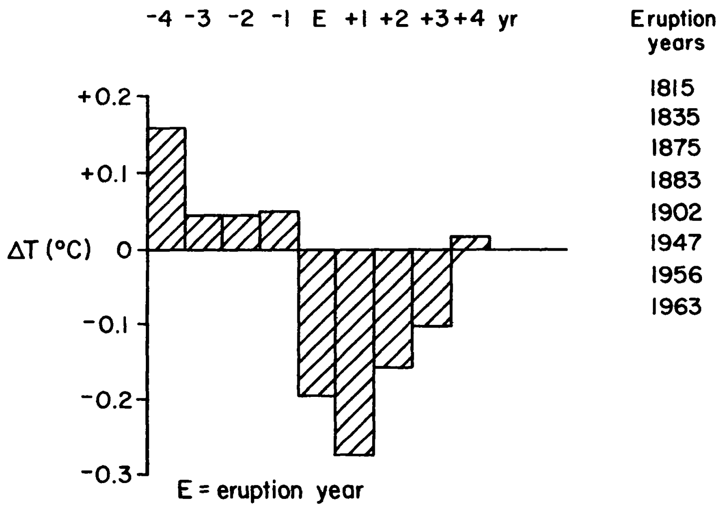

Figure 1 Composite plot of the temperature departure for the Northern Hemisphere in the four years immediately before and after some large nineteenth and twentieth century eruptions (after Self et al 1981).

volcanic perturbations and to reexamine them in detail. Hansen et al (1978) examined the 1963 eruption of Mt. Agung on the Indonesian island of Bali, which occurred at a time when tropospheric and stratospheric temperatures were being routinely measured (Newell 1970, Newell & Weare 1976) and accurate measurements of aerosol optical depth were being made at a number of observatories (Volz 1970). [Optical depth is equal to the negative natural logarithm of the attenuation of incident light, or $\tau = -\ln(I/I_0)$, where I_0 and I are the initial and final light intensity, respectively.] High-altitude aircraft above 20 km could also directly collect stratospheric aerosols for analysis (Mossop 1964).

Hansen et al (1978) calculated the expected effect on the stratospheric and tropospheric temperatures by using the measured time history of aerosol optical depth and a simple one-dimensional radiative/convective climate model. They then compared these calculated temperature changes with the observed temperature perturbations. The theoretical results agreed with the observation that stratospheric temperatures rose by 4 to 8°C in the region from 10°N to 30°S, while surface temperatures in the region from 30°N to 30°S showed a decrease of a few tenths of a degree over a period on the order of a year. Hansen et al (1981) later confirmed, to a large extent, their earlier theoretical results for surface cooling as well as the simpler box-model predictions of Schneider & Mass (1975) and Harshvardhan & Cess (1976). Further studies of the effects of volcanic aerosols on climate using more sophisticated two-dimensional climate models have in general confirmed the empirical evidence for cooling of a few tenths of a degree Celsius following large historic eruptions (Robock 1981, 1984, Chou et al 1984, McCracken & Luther 1984).

A major finding in both the statistical and individual studies of volcanic perturbations of the atmosphere is that relatively small volcanic eruptions (measured by the total volume of magma ejected as pumice and ash) such as that of Mt. Agung in 1963 [estimated volume of ejected magma of only 0.3–0.6 km^3, dense-rock equivalent (DRE); Rampino & Self 1984a] can lead to a relatively dense aerosol cloud, totaling perhaps 10 megatons (Mt) of H_2SO_4 aerosols in the specific case of Agung [at least in the Southern Hemisphere; the Northern Hemisphere stratosphere contained only about one fifth of this mass of aerosols (see Table 1)]. By contrast, the much larger Krakatau eruption in 1883 [$\sim$10 km^3 (DRE) of ejected magma] produced a cloud only five times more massive ($\sim$50 Mt of H_2SO_4 aerosols). Obviously a number of factors, both volcanological and meteorological, involving the amount of H_2SO_4 aerosols created in the upper atmosphere and their potential for widespread distribution come into play in determining the impact of any particular volcanic eruption on the climate.

Volcanologists have attempted to measure the amount of sulfur volatiles emitted by past eruptions through analysis of the composition of the solid products (pumice, scoria, lava, or fine ash). One method is to determine the sulfur volatile content in glass inclusions in crystals of minerals that formed within the magma chamber just prior to the eruption; this gives a

Table 1 Estimates of stratospheric aerosols and climatic effects of some volcanic eruptions[a]

Volcano	Latitude	Date	Stratospheric aerosols (Mt)	Northern Hemisphere τ_D	Northern Hemisphere ΔT (°C)
Explosive eruptions					
St. Helens	46°N	May 1980	0.3	<0.01	<0.1
Agung	8°S	March/May 1963	10	<0.05[b]	0.3
El Chichón	17°N	March/April 1982	20	0.15	<0.4
Krakatau	6°S	August 1883	50	0.55	0.3
Tambora	8°S	April 1815	200	1.3	0.5
Rabaul?	4°S	March 536	300	2.5	large?
Toba	3°N	−75,000 yr	1000?	10?	large?
Effusive eruptions					
Laki	64°N	June 1783 to February 1784	~0	Locally high[c]	1.0?
Roza	47°N	−14 Myr	6000?	80?[d]	large?

[a] References: Rampino & Self (1984a) and text. Optical depths are visual, direct beam.
[b] Southern Hemisphere $\tau_D \approx 0.2$.
[c] Aerosols were mostly tropospheric.
[d] If the aerosols were dispersed globally, the average Northern Hemisphere optical depth would have been about 40.

measure of preeruption sulfur. Knowing the volume of magma erupted, one can then estimate the sulfur volatile release by determining the amount of sulfur contained in the erupted ash and taking the difference (Sigurdsson 1982, Devine et al 1984). These methods are currently being refined (Sigurdsson et al 1985).

Since a positive general correlation has been established between the solubility of sulfur and the iron content in a magma, basaltic (mafic) magmas tend to be richer in dissolved sulfur than more silicic magmas. Sulfur release into the atmosphere from a basaltic eruption may be an order of magnitude greater than that of a silicic eruption of similar volume, but this is balanced in climatic impact by the fact that silicic eruptions in general are more explosive and therefore tend to create eruption columns that reach well into the stratosphere (Wilson et al 1978). Some eruptions, however, may be "anomalously" rich in sulfur volatiles with regard to their major element chemistry, such as was the case for the 1982 eruption of El Chichón (Table 1), which is discussed later.

Analyses of volcanic emissions show that sulfur is emitted mostly as sulfur dioxide (SO_2) and also as hydrogen sulfide (H_2S), which is soon oxidized to SO_2. In the stratosphere, the sulfur dioxide reacts with hydroxyl (OH^-) radicals produced by the photodissociation of water vapor. Gaseous sulfuric acid condenses on minute seed particles of dust (possibly volcanic or meteoritic) or on ions or small clusters of molecules. The photochemical reactions may be protracted, with complete conversion of emitted sulfur gases into aerosols taking weeks to months.

The residence time of the aerosols depends upon the dynamics of nucleation and growth of the droplets. After the initial input of sulfur volatiles and the conversion to droplets, the volcanic aerosols typically have a modal diameter of about half a micron [see Turco et al (1982) for a comprehensive review of observations and theory related to stratospheric aerosols]. For historic eruptions of all sizes, from those just capable of stratospheric injection (e.g. Fuego in 1974) to the largest known (e.g. Tambora in 1815), the *e*-folding time for fallout of the stratospheric aerosols has been observed to be about one year (Stothers 1984a). This means that for significant eruptions (those that create more than 1 Mt of stratospheric sulfuric acid aerosols) the stratospheric aerosol optical depth can be perturbed for several years.

INFORMATION FROM ICE CORES

Reasonably continuous, direct estimates of atmospheric optical depth from astronomical observations of the Sun, Moon, and stars are available only for the time period since 1883. During this period, changes in the atmospheric transparency have been correlated with significant volcanic eruptions (e.g. Pollack et al 1976). A major problem is to establish methods of estimating the amounts of sulfur aerosols created by significant eruptions prior to that time. In an important paper, Hammer et al (1980) presented evidence from the yearly ice layers in deep Greenland ice cores for sharp increases in acidity that coincided with the times of historic eruptions. They used the acid concentrations to estimate the global stratospheric aerosol burden in these volcanic years.

The ice-core method has several limitations, however, including the fact that eruptions in relatively high northern latitudes can produce especially large acidity spikes because of closeness of the volcanoes to Greenland. For example, the years 1963 and 1964 are associated with a noticeably high acidity peak in Greenland ice cores—enough to suggest about 20 Mt of aerosols in the Northern Hemisphere stratosphere if the source were the equatorial Agung eruption. But we know from direct atmospheric observations that less than one fifth of this amount was actually spread in the stratosphere north of the equator from the Agung eruptions in Bali (Volz 1970). The Greenland acidity spike is too high to be the result of Agung aerosols and is almost certainly due to tropospheric transport from smaller nearby eruptions such as the ongoing Surtsey eruptions off Iceland (Cronin 1971). Significantly, Koerner & Fisher (1982) detected no excess ice acidity for the years 1963 and 1964 on Ellesmere Island, Canada, whereas Delmas & Boutron (1980) and Legrand & Delmas (1987) did discern a strong acid signal in Antarctic ice which could be attributed to Agung. Icelandic eruptions in general are overrepresented in Greenland ice cores (Hammer 1984). Globally, however, the effectiveness of transport of acid aerosols to Greenland may also vary with seasonal and year-to-year changes in atmospheric circulation patterns (Hammer et al 1980).

It is worth emphasizing that if an eruption, even a moderately large one, does not inject sufficient sulfur into the atmosphere it will not appear above the noise level in the polar ice acidity record. This means that most eruptions with potential climatic impact are expected to leave a discernible trace in polar ice, although the amount deposited may not be proportional to the actual mass of sulfur produced, for the meteorological reasons just discussed. Probably any explosive eruption bigger than Krakatau's in 1883 can be detected in one or both of the polar ice sheets, because so much magma is erupted that the sulfur release is bound to be fairly large in any case.

One way of correcting for all these problems is by making comparisons with ice cores from several other localities to get a better estimate of the global distribution of the aerosols. Although such a method of calibration holds promise, for example, with the conductivity records of the Quelccaya ice core from the Peruvian Andes (Thompson et al 1986) and the Yukon ice core (Holdsworth et al 1986), it has proven to be difficult even with cores from different parts of the Greenland Ice Sheet (Herron 1982). The accumulation rate of snow in Antarctica is much less than that in Greenland, and hence the yearly ice layers there are generally thinner and more difficult to count and date accurately (Delmas et al 1985). Obviously, the best way of calibrating the ice-core acidity records as a record of global stratospheric aerosols is to identify, if possible, the location of the specific eruptions that produced the acid spikes.

It is worth noting that some periods of enhanced ice-core acidity and microparticle accumulation, lasting for decades and longer, coincide with historic cool intervals such as the Little Ice Age (Hammer et al 1980, Porter 1981, 1986, Thompson & Mosley-Thompson 1981, Thompson et al 1986). It is possible, therefore, that episodes of greater-than-average volcanism may modulate the climate over periods of tens to hundreds of years (for a review, see Bryson & Goodman 1980).

MT. ST. HELENS AND EL CHICHÓN: A STUDY IN CONTRASTS

The spectacular Mt. St. Helens eruption of May 1980 produced a stratospheric cloud of ash but released a relatively small amount of SO_2 into the Northern Hemisphere stratosphere, with the result that only about 0.3 Mt of sulfuric acid aerosols were produced (Table 1) and no significant climatic effects [aside from a local daytime cooling the next day for stations downwind of the ash (Robock & Mass 1982)] were detected (Newell & Deepak 1982). In March–April 1982, however, the Mexican volcano El Chichón erupted explosively and sent a huge cloud rich in SO_2 up to about 26 km in the stratosphere. Observations from the ground and by satellite showed that this eruption was having a large impact on the stratosphere, and the spread of the cloud could be tracked accurately.

El Chichón provided the test case for which volcanologists and atmospheric scientists had been waiting (see reviews by Rampino & Self 1984b, Hofmann 1987). Like St. Helens, it was a small eruption volumetrically, producing only about 0.3 to 0.4 km^3 (DRE) of magma, but it was extremely rich in sulfur (derived perhaps from deposits of $CaSO_4$ beneath the volcano); thus, while the volcanic ash contribution to the atmosphere was small, the sulfuric acid aerosol contribution was considerable, about 20 Mt (Table 1). This is enough, theoretically, to lower the surface temperatures in the Northern Hemisphere a few tenths of a degree Celsius, and the year 1982 showed such a cooling (Table 1), although it seems that the cooling began with colder than average weather from January to March—before the El Chichón eruptions. But the summer was cool, with a unique snowfall in Vermont in August. Kelly & Sear (1984) proposed that Northern Hemisphere eruptions can cause cooling within the first 2 to 3 months after an eruption. Longer term effects from El Chichón were predicted by some models (Robock 1984), and Reiter & Jäger (1986) suggested that the cold winter of 1984/1985 was possibly related to lingering aerosols from the eruption.

The severe El Niño event of 1982/1983 added some confusion as to climatic cause and effect. Handler (1984), among others, has suggested that the El Chichón aerosol cloud either triggered the El Niño or led to its intensification—a proposal that has generated a good deal of debate. (It is interesting to note that the 1963 Agung eruption was also followed by an "off-season" El Niño event.)

THE GREATEST HISTORIC ERUPTIONS AND THEIR ATMOSPHERIC EFFECTS

With the Greenland ice-core record of acidity as a guide to notable "eruption years" (Hammer et al 1980, Hammer 1984), it has become possible to attempt to identify the source volcanic eruptions for the acid-rich layers deposited on the polar ice sheets. For more recent historical times it is relatively easy to pinpoint the eruptions that caused acid spikes (for example, Krakatau in 1883 and Tambora in 1815), but even here the situation may be more complicated than it first appears. Mt. Augustine in Alaska also erupted in 1883, and the Mayon eruption in the Philippines in 1814 may have contributed some acids to the ice layers of 1815 and 1816 (Stothers 1984a). Asama in Japan erupted at the same time as Laki during 1783.

In order to identify the extent of aerosol clouds and possible sources of ice-core acidity spikes prior to AD 1500, it has been necessary to search through historical records for evidence both of local eruptions (mostly in the Mediterranean region) and of the atmospheric perturbations caused by aerosol clouds from perhaps distant eruptions. A virtually complete search of the European records prior to AD 630 (Stothers & Rampino 1983a,b) turned up occasional evidence of significant atmospheric disturbances, such as a dim Sun and Moon, unusual atmospheric optical phenomena, and unusually cold weather accompanied by crop failures and famine. Similar work is now being done for the extensive Chinese historical records (Pang & Chou 1984, 1985, Pang et al 1986). It will be useful in this review to summarize the atmospheric and climatic effects occurring in some of the most severe and better known of the historical eruption years.

1816: The Tambora Effect and the Year Without a Summer

The year 1816 has gone down in the annals of climate history as the "Year Without a Summer" and "Eighteen Hundred and Froze to Death" (Stommell & Stommell 1983). In fact the entire decade from 1810 to 1820 was a time of noticeably cool temperatures in the Northern Hemisphere, and this has been correlated by some authors with the low sunspot maximum in 1816 (e.g. Humphreys 1940). The unusual weather in 1816 followed the spectacular April 1815 eruption of Tambora volcano on Sumbawa Island in Indonesia—one of the largest known ash-producing eruptions [150 km^3 of ash and pumice, equal to about 50 km^3 (DRE) of magma] in the last 10,000 years (Stothers 1984a, Self et al 1984).

Ash fallout was noted over an area in excess of 4×10^5 km^2 (and probably fell over an area of more than 10^6 km^2); darkness lasted for up to 2 days at distances of 600 km from the volcano. The eruption rate and the area of ash dispersal both suggest that the eruption column may have reached 50 km into the stratosphere. The volcanic cloud traveled around the world, and within 3 months its optical effects were observed at distant locations in Europe. For example, around the end of June, and later in September, several observers near London reported prolonged and brilliantly colored sunsets and twilights.

The following year (1816) was marked by a persistent dry fog, or dim Sun, as reported in the northeastern United States. The haze was clearly located above the troposphere, since neither surface winds nor rain dispersed it and because the total lunar eclipse of 9–10 June was extremely dark. Stothers (1984a) has derived a time history of the optical properties of the Tambora aerosol cloud (Figure 2) by using indicators of reduced atmospheric transmissivity such as dimming of the Sun (shown by increased naked-eye visibility of sunspots) and dimming of starlight (noted by astronomical observers). The calculated aerosol mass for Tambora (Table 1) is in good agreement with estimates based on the 4-yr-long (1815–1818) acidity enhancement in the Crête, Greenland, ice core (Hammer et al 1980). Fallout from this eruption has probably also been detected in Antarctic ice (Thompson & Mosley-Thompson 1981, Delmas et al 1985, Legrand & Delmas 1987).

The exceptional meteorological conditions spawned by the explosion started with a hot, followed by an "extremely cold," pocket of air directly under the tropospheric ash clouds (at least at Banjuwangi, 400 km from the volcano) and then continued with freezing temperatures in Madras, India, two weeks later (Stothers 1984a). Analogous studies of the effects of the Canadian wildfires of 1950 (Wexler 1950) and the Siberian wildfires of 1915 (Seitz 1986) have shown that surface temperatures in those cases

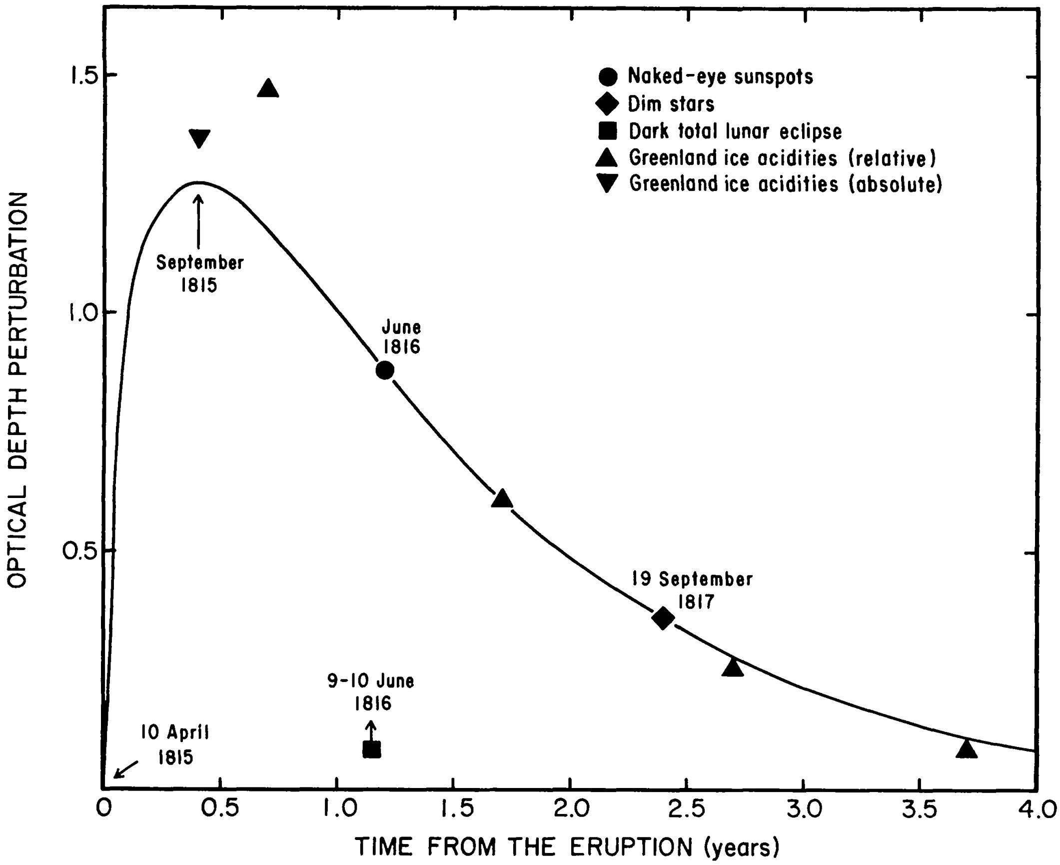

Figure 2 Excess visual optical depth at northern latitudes as a function of time reckoned from the date of the 1815 Tambora eruption. The plotted point for 9–10 June 1816 is only the lower limit to the true value (after Stothers 1984a).

dropped several degrees Celsius in areas that were thickly covered by high-altitude smoke clouds, as a result of the attenuation of the incoming solar flux. On the other hand, low-altitude smoke clouds heated up the boundary layer locally. Since volcanic ash sufficiently resembles sooty smoke, the meteorological analogies probably can be validly made in a qualitative way.

The summer of 1816 in western Europe was cool and exceedingly wet; crop failures (compounded by the aftereffects of the Napoleonic Wars) led to famine, disease, and social unrest, referred to by Post (1977) as "The Last Great Subsistence Crisis in the Western World." Kelly et al (1984) have suggested that an important effect of volcanic aerosol clouds is to produce a marked drop in surface pressure across the midlatitudes of the North Atlantic sector, leading to a southward shift in the track taken by middle-latitude cyclones. A major anomaly would thus be centered over England and would extend over much of western Europe, giving rise to a cold, wet summer. In support of this, Kelly et al have reconstructed pressure anomaly charts of Europe based on the available data; these charts are dominated by negative pressure anomalies over Europe beginning in early 1816. Data from Manley (1974) show that the summer months of 1816 in central England were about 1.5°C cooler than during the summer of 1815. The dismal European summer is credited with having inspired Mary Shelley to write *Frankenstein*, and Lord Byron his poem *Darkness*.

In North America, records of Hudson's Bay Company posts on the eastern side of Hudson Bay show that the summers of 1816 and 1817 were the coldest of any in the modern record (Wilson 1985a,b). Tree-ring data from northern and western Quebec support these observations (Filion et al 1986, Jacoby et al 1987). The distribution and severity of sea ice in Hudson Strait in 1816 suggests prevailing northerly or northwesterly winds, which again supports the idea that these years were marked by the development of strongly meridional atmospheric circulation patterns allowing southward penetrations of Arctic air across eastern North America and western Europe (Catchpole & Faurer 1983). Outbreaks of unusually cold weather during the spring and summer of 1816 in eastern Canada and the eastern United States are well documented (Post 1977, Stommel & Stommel 1983, Hamilton 1986). For example, the summer of 1816 was the coldest in New Haven, Connecticut, for the entire period from 1780 to 1968 (Landsberg & Albert 1974). From late spring through the summer, repeated frosts in New England caused crop failures, resulting in poor harvests and food shortages.

The outbreaks of cold weather and raininess during the summer months of 1816 are seen clearly in a number of other climatic indicators from around the world, from lateness in the grape harvests in France (Stommel & Swallow 1983) to frost damage rings in trees in the western United States (LaMarche & Hirschboeck 1984) and in South Africa (Dunwiddie & LaMarche 1980).

On a zonal to hemispheric basis, the deviation of annual mean temperature is more difficult to assess, since the station coverage in 1816 was very spotty. Using W. Köppen's compilation of temperature data, Stothers (1984a) finds an average deviation for the "Northern Hemisphere" in 1816 of -0.4 to -0.7°C, whereas the value for northern midlatitudes is about -1.0°C. This agrees with other estimates based on less data and somewhat different averaging (Lamb 1970, Rampino & Self 1982, Angell & Korschover 1985).

1783: The Fire Fountains of the Laki Fissure Eruption

Franklin's observations of the "dry fog" produced by the Laki (or Lakagígar) eruption have focused the attention of climatologists and volcanologists on the events surrounding this unusual eruption and its aftermath. The Laki eruption began in June 1783 and lasted for 8 months. The eruptions were not the typical explosive eruptions of the sort that produce great amounts of pumice and ash; Laki was primarily a fissure basalt, lava-flow type, and it erupted about 12.3 km^3 of lava, the bulk of it coming from a 13 km length of fissure during June and July, in the first 50 days of the eruption (Thorarinsson 1969). From eyewitness accounts, it appears that during the first days the eruption was extremely violent, with "enormous" Hawaiian-type lava fountains. Thorarinsson (1969) estimated that about 0.3 km^3 (DRE) of tephra was erupted, mostly during this early violent phase, and fine ash from the eruption fell as far away as northern Europe.

The effects of the eruption in Iceland were disastrous. The toxic volcanic gases and aerosols created a "blue haze" that spread all across the island and led to the destruction of the summer crops. About 75% of the livestock in Iceland died, and the resulting "Blue Haze Famine" claimed 24% of the Icelandic population. The dry fog reported by Franklin was also reported by others in Europe (for example, Gilbert White in England) and was even seen in Asia and North Africa (Holm 1784, Russell & Archibald 1888). Wood (1984) has established from eyewitness reports that much of the haze over Europe lay in the lower troposphere. However, the reported visibility of the haze high up in the Alps and its continued observability throughout Europe for weeks in spite of changing wind directions and rainfall suggest that it extended upward at least into the upper troposphere. The haze in Europe appeared most intense during June and July, precisely the same months that Laki was most active. The eruption ceased by early February 1784, but the dry fog had already largely disappeared by the end of December 1783 (Stothers et al 1986). In three Greenland ice cores, Hammer et al (1980) and Hammer (1984) found the acidity of the 1783 layers to be extraordinarily high, but no excess acidity was found in the 1784 layers, contrary to what one would expect if the stratosphere had been significantly loaded with aerosols. Moreover, the total lunar eclipse of 10 September 1783 was not unusually dark (Maclean 1984). Thus, the eruption column of fine ash and volcanic gases from Laki must have normally reached only up to, at most, the tropopause (8 to 11 km during the Icelandic summer).

In Iceland itself, a prompt and extreme cooling at the surface was observed directly under the ash clouds during the summer of 1783 (Stephensen 1813). Elsewhere, as Sigurdsson (1982) has shown from early Northern Hemisphere temperature records, an abnormal temperature decline began in the autumn of 1783 and reached a minimum in the period from December 1783 to February 1784. This period showed the lowest mean winter temperature in 225 years, which was 4.8°C below the long-term average. The spring, autumn, and winter mean temperatures for 1784 and 1785 were below normal as well. Northern tree-ring data also indicate cold growing seasons during 1783 and 1784 (Oswalt 1957, Filion et al 1986). Theoretically, very fine solid particles and *small* tropospheric aerosols in continuous production, with extensive horizontal dispersion, might have been able to initiate such a cooling (see Hansen et al 1980).

536: The Mystery Cloud

The densest and most persistent dry fog in recorded history was observed during AD 536–537 in Europe and the Middle East (Stothers & Rampino 1983a, Stothers 1984b) and in China (K. D. Pang & H.-h. Chou, in Weisburd 1985). In the Western literature five reliable contemporary descriptions of the atmospheric conditions of 536–537 exist. According to one contemporary writer (probably John of Ephesus), conditions in Mesopotamia (30° to 37°N) were such that "the sun was dark and its darkness lasted for eighteen months; each day it shone for about four hours, and still this light was only a feeble shadow . . . the fruits did not ripen and the wine tasted like sour grapes." The winter in Mesopotamia was exceptionally severe, with freak snowfalls and much hardship. In Italy, a high government official (Cassiodorus Senator, not mentioned by Stothers & Rampino) wrote in the late summer of 536: "The sun . . . seems to have lost its wonted light, and appears of a bluish color. We marvel to see no shadows of our bodies at noon, to feel the mighty vigor of the sun's heat wasted into feebleness, and the phenomena which accompany a transitory eclipse prolonged through almost a whole year. The moon, too, even when its orb is full, is empty of its natural splendor. . . . We have had . . . a spring without mildness and a summer without heat . . . the months which should have been maturing the crops have been chilled by north winds . . . rain is denied . . . the reaper fears new frosts" (Cassiodorus Senator AD 536). Cold and drought finally succeeded in killing off the crops in Italy and Mesopotamia and led to a terrible famine in the immediately following years. These and other accounts of the time read remarkably like Franklin's and others' modern descriptions of the "dim Sun" conditions following known volcanic eruptions.

It is possible to estimate from these historical accounts that such a pronounced reduction in solar brightness would require an excess visual atmospheric optical depth of $\tau_D = 2.5$ (Stothers 1984b). Under these conditions, at maximum altitude, the Sun (and Moon) would have appeared about 10 times fainter than normal, thus accounting for the reported darkening; scattered sunlight would have illuminated the rest of the sky. The mystery cloud first appeared in the Mediterranean region in late March of 536. Observers at 41 to 42°N reported effects lasting for 12 to

15 months, whereas at 30 to 37°N the duration recorded was 18 months. This suggests that the eruption that produced the aerosols was situated somewhere to the south; a possible source is the large eruption of Rabaul (4°S), on the island of New Britain off New Guinea, radiocarbon dated to AD 540±90 (Heming 1974).

Similar atmospheric effects were seen in China during the same years. Pang & Chou (see Weisburd 1985) have recently noted, for example, that the bright star Canopus was not visible when looked for at the equinoxes of 536. They have also documented and reconstructed the distribution of summer snows and frosts, drought, and famine throughout China in the years 536–538. The situation in China clearly paralleled that of Europe and the Middle East; the mystery cloud and the anomalously cold weather seem to have occurred throughout at least a large portion of the Northern Hemisphere. Moreover, a deep Greenland ice core shows a high acidity at roughly this date, originally given as 540±10 (Hammer et al 1980, Herron 1982), but later revised to 516±4 (Hammer 1984) and perhaps still in need of revision.

44 BC: Caesar's Death and the Year of the Failing Sun

Another significant dimming of the Sun is reported in the ancient literature for 44 BC and the subsequent two years. In the Western records (Stothers & Rampino 1983a), Plutarch, writing ca. AD 100, gives the fullest account of the "dim Sun" conditions after the murder of Julius Caesar: "For during all that year its orb rose pale and without radiance, while the heat that came down from it was slight and ineffectual, so that the air in its circulation was dark and heavy owing to the feebleness of the warmth that penetrated it, and the fruits, imperfect and half ripe, withered away and shrivelled up on account of the coldness of the atmosphere." Ovid (ca. AD 10) describes the Moon as "bloody" and Venus as "darkly rusty" in 44 BC, while Calpurnius Siculus (ca. AD 60) alludes to the "bloody" color of the comet of 44 BC (not cited by Stothers & Rampino). One possible cause for these atmospheric conditions was an eruption of Etna dated to the same year. As Vergil relates, "Mighty Etna . . . from its burst furnaces breathes forth flame; and . . . all Sicily moans and trembles, veiling the sky in smoke." Livy and, later, Pliny the Elder independently testified to the exceptional magnitude of this explosive Etnan eruption. Etna's last large eruption had occurred about 77 years earlier. (We now think that Lucan's and Petronius's apparent allusions to an eruption in 50–49 BC actually refer to the 44 BC eruption.)

Atmospheric effects in China were reported in the *Chronicles of the Han Dynasty* (Schove 1951, Pang & Chou 1984). For example, in April–May 43 BC, "It snowed. Frosts killed mulberries." In May–June, "The sun was bluish white and cast no shadow. At high noon there were shadows but dim." In October, "Frosts killed crops, widespread famine. Wheat crops damaged, no harvest in autumn." The historical data and the Greenland ice-core acidity record support the idea of multiple eruption clouds in the years from 44 to 42 BC (Stothers & Rampino 1983a, Pang et al 1986). A strong 3-yr-long acidity peak in two Greenland ice cores has been dated around 50 ± 4 BC (Hammer et al 1980, Hammer 1984; see also Herron 1982), and it probably correlates with the veiled Sun and other peculiar optical phenomena of 44–42 BC.

COMETARY WINTER AND NUCLEAR WINTER

Interest in the aftermath of a proposed collision of an asteroid or comet with the Earth at the time of the Cretaceous-Tertiary mass extinctions (66 Myr BP) has led to scenarios of Sun-blocking dust clouds originating from the huge cratering event, and smoke clouds rising from widespread wildfires [see Alvarez (1986) for a review]. These studies created the impetus for an analysis of the possible atmospheric effects of the sooty smoke from burning cities in the wake of a nuclear war (Crutzen & Birks 1982, Turco et al 1983). The initial "nuclear winter" simulation studies, based on the results of simple one-dimensional radiative/convective climate models, suggested the possibility of drastic temperature decreases of up to 30°C, with subfreezing conditions for weeks to months over large portions of the Northern Hemisphere and effects penetrating into the Southern Hemisphere as well, after a "baseline" nuclear exchange (Turco et al 1983, 1984b). This work prompted several research groups to study the effects of smoke generated by nuclear war, and the results obtained touched off a debate regarding the uncertainties in the amount of smoke that could be lofted to high altitudes and the scale and severity of the climatic effects of nuclear wildfires (for a review, see Colbeck & Harrison 1986).

Two independent study groups (National Research Council 1985, SCOPE 1985) concluded that the climatic effects could be severe, but that there were many uncertainties in the nuclear winter analyses. The most recent studies, using more sophisticated climate models (GCMs), suggest that the cooling would be much less severe though still significant ("nuclear autumn"), with worst case (July) decreases of perhaps 5°C in low latitudes (10–30°N) and 10–15°C at higher northern latitudes, lasting only a few weeks and with considerable unevenness. Most of the reduction in the degree of cooling is related to the probable removal of 75% of the smoke from the atmosphere within the first 30 days, as well as to patchiness in the smoke distribution and to the moderating effects of the oceans in the climate system (Thompson & Schneider 1986, Covey 1987).

VOLCANIC WINTER?

Large Explosive Eruptions

We have thus far discussed the climatic aftereffects of a number of the greatest historical volcanic eruptions. The nuclear winter debate raised a suggestion that the atmospheric aftereffects of these volcanic eruptions might be used as a basis for estimating the severity of cooling from dense smoke clouds (Maddox 1984, Brown & Peczkis 1984). The differing optical properties between volcanic aerosols and black, sooty smoke from urban fires, however, makes such a comparison difficult (Turco et al 1984a, 1985). More importantly, historic eruptions have produced relatively small amounts of aerosols. But perhaps the historic eruptions can be used as a baseline for estimating the possible atmospheric effects of the largest volcanic eruptions in the geologic past—much larger than recent historic events such as Tambora or Krakatau, or even the mystery eruption of 536. One good example is the Toba eruption in Sumatra about 75,000 years ago, which is the best-known late Quaternary "supereruption," with a recent estimate of the volume of erupted pyroclastics being equivalent to more than 2000 km^3 (DRE) of magma (Rose & Chesner 1986).

The Toba ash layer is extraordinarily widespread (Ninkovich et al 1978), and rough calculations, using the same methods as Murrow et al (1980), suggest that a maximum of $\sim 0.8\%$ of the erupted material could be in the form of fine dust $< 2\ \mu m$ in diameter, for a total of about 20,000 Mt of volcanic dust. If only 10% of this dust were injected into the stratosphere, conditions of total darkness could have existed over a large area for weeks to months (see also Kent 1981).

In order to estimate any longer-lasting atmospheric effects, however, it is necessary to calculate the total amount of sulfuric acid aerosols that could have been produced by the Toba eruption. From simply scaling upward from historical eruptions of similar composition magma, Toba is estimated to have been capable of producing between 1000 to 5000 Mt of sulfuric acid aerosols (Rampino et al 1985). For volcanic aerosols, the globally averaged optical depth is $\tau_D = 6.5 \times 10^{-3}\ M_D$, where M_D is the global aerosol loading in megatons (Stothers 1984a). For Toba the equivalent global aerosol optical depths are 6 to 33 (Figure 3). Local, regional, and hemispheric effects could have been greater, depending on the spread of the cloud. These values may be compared with the peak aerosol optical depth of about 2 estimated for the AD 536 mystery cloud or the value of about 1 following the 1815 Tambora eruption. The atmospheric aftereffects of a Toba-sized explosive eruption might be comparable to some scenarios of nuclear winter, although the aerosols are expected to have a longer atmospheric residence time than would the nuclear winter smoke.

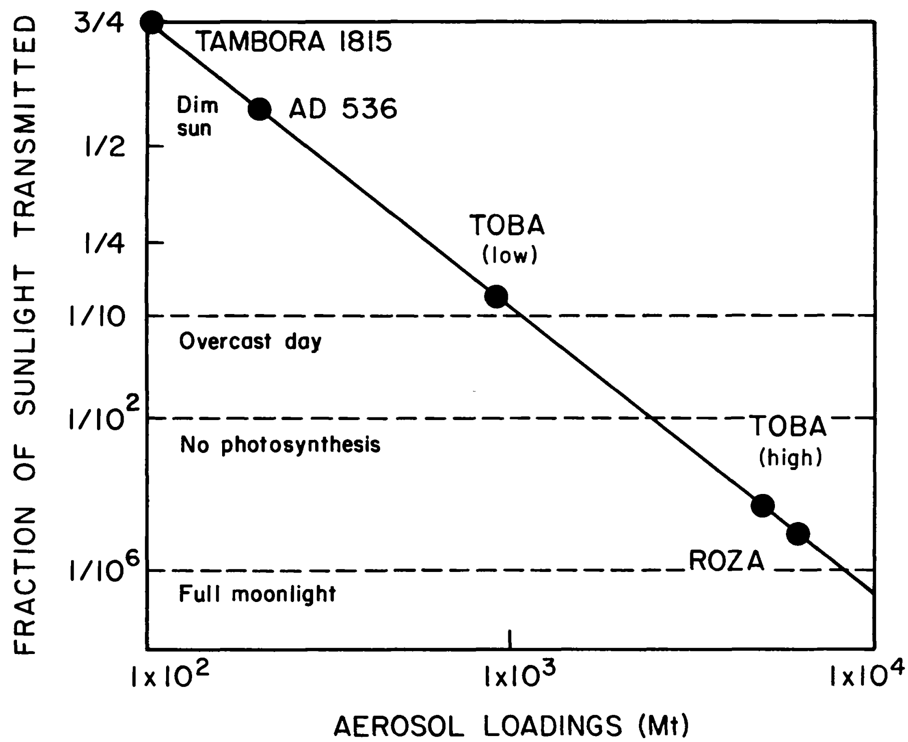

Figure 3 Fraction of sunlight transmitted through stratospheric aerosol and/or fine-ash dust clouds of different masses (theoretical line, after Turco et al 1984b). Points refer to great historic and prehistoric eruptions (see text).

If such an aerosol cloud could form and persist in the stratosphere, the climatic effects would almost certainly be quite severe. It is important to stress, however, that these are "worst-case" situations, made simply by extrapolating a linear increase in mass of aerosols under the assumption that the behavior of very dense aerosol clouds is not qualitatively different from that of the less dense aerosol clouds observed after historical eruptions. Recent work has shown that differences in aerosol nucleation, saturation, and fallout in dense clouds may affect the concentrations and atmospheric lifetimes of the aerosols. Pinto et al (1987) have recently used a one-dimensional radiative/convective model including aerosol microphysical and photochemical processes to examine the conversion of sulfur dioxide to aerosols in the stratosphere after volcanic eruptions. They find that for successively larger injections of SO_2, in the range of 10 to 200 Mt, the processes of condensation and coagulation produce larger particles; these particles have a smaller optical depth and fall out of the stratosphere faster. The rate of SO_2 oxidation may also be limited by conversion of OH to HO_2 radicals, which could limit the formation of aerosols. These results all suggest that the buildup of H_2SO_4 aerosols in the stratosphere might be self-limiting to a degree. However, Pinto et al have not yet modeled the injection of SO_2 burdens of >1000 Mt, accompanied by possibly large amounts of water vapor, as could be the case after "supereruptions" such as Toba. In such cases the dynamics of gas to particle conversion and the *e*-folding time of aerosols and ash in the stratosphere and troposphere might be quite different from those in the less dense clouds that have been observed and modeled thus far.

Flood-Basalt Eruptions

As mentioned above, basaltic volcanic eruptions may release an order of magnitude more sulfur volatiles than do silicic eruptions of the same volume (Devine et al 1984). Very recent results indicate that episodes of flood-basalt volcanism in the geologic past have involved the outpouring of up to 10^6 km^3 of basaltic magma over peak time periods of less than a million to a few million years (Bellieni et al 1984, Courtillot & Cisowski 1987, White 1987). Individual eruptions seem to have generated tens to hundreds of cubic kilometers of magma in periods of days to weeks. In the past, "quiet" effusive basaltic eruptions were considered unlikely to produce high-altitude aerosol clouds (Lamb 1970). Recent study has shown, however, that even relatively small historic fissure basalt eruptions, such as the 1783 Laki eruption, have produced widespread aerosol clouds. Theoretical plume modeling of such eruptions indicates that at rapid eruption rates the sulfur volatiles are efficiently released and can be carried to high altitudes in convective plumes rising above large fire fountains (Figure 4).

Stothers et al (1986) have recently considered the possible atmospheric impact of the large flood-basalt eruptions in the geologic past. For example, the Roza flow eruption of the Columbia River Basalt Group (about 14 Myr BP) produced some 700 km^3 of basaltic lava in about 7 days (Swanson et al 1975). The estimated eruption rates of 10^4–10^5 m^3 s^{-1} from 1 to 10 km fissure lengths are predicted to have generated Hawaiian-type fire fountains about 1 km in height (Wilson & Head 1981) and stratospheric (>10 km) eruption plumes (Figure 4).

The quantity of atmospheric aerosols produced by such large basalt eruptions can be roughly estimated by scaling from the known amounts of aerosols generated by the largest modern fissure eruptions, such as Laki in 1783. Laki erupted about 12 km^3 of magma and is estimated by various methods to have released about 30 Mt of sulfur (Stothers et al 1986). The sulfur release from the Roza flow eruption is therefore computed to have been (700 km^3/12 km^3) $\times$ 30 Mt $\approx$ 2000 Mt, equivalent to about 6000 Mt of H_2SO_4 aerosols; the corresponding global average optical depth would be about 40. Such a thick aerosol cloud, distributed worldwide, would allow only a small fraction (10^{-5}) of sunlight to reach the Earth's surface

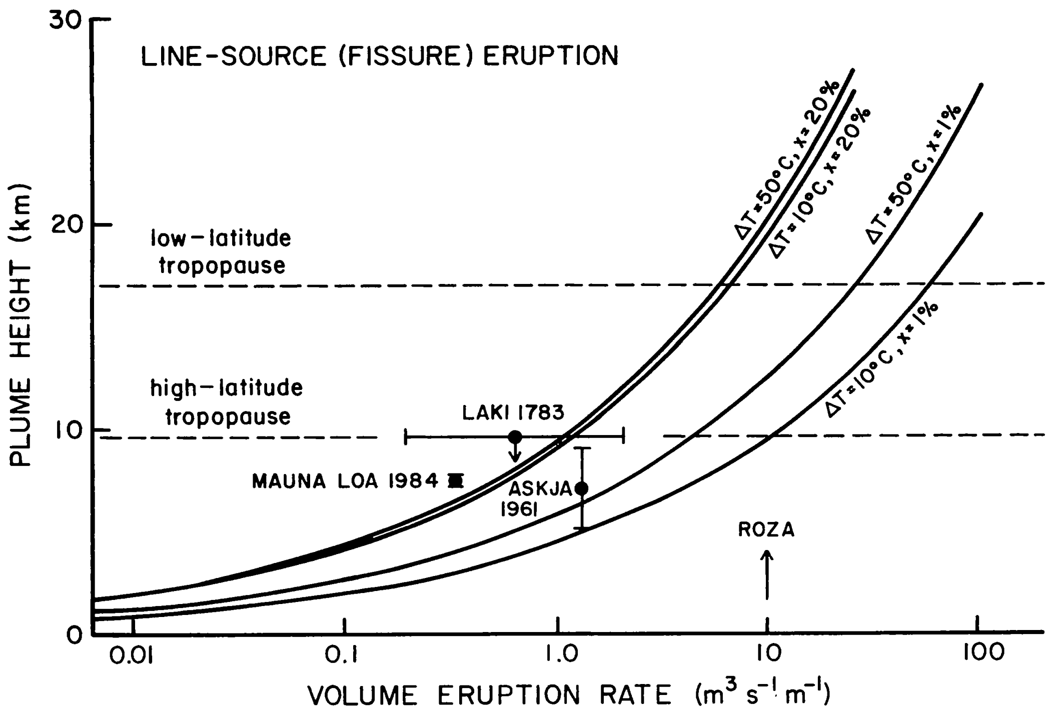

Figure 4 Convective plume height as a function of volume eruption rate per meter length of fissure for a line-source eruption. Theoretical curves are given for two values of the fine-ash content (x) and for two values of the temperature drop of the fountain clasts (ΔT) appropriate for fire-fountain activity. Predicted plume heights for the Roza flow eruption can be read off the observationally calibrated theoretical curves. The plotted plume height for Laki is an observational upper limit (after Stothers et al 1986).

(Figure 3). In this case, barring efficient self-limiting mechanisms, the cloud's atmospheric effects would be comparable to those in recent nuclear winter models, but more extended in time.

VOLCANISM AND MASS EXTINCTION?

One of the great current debates in geology concerns the cause of mass extinctions. Strong evidence now exists for a comet or asteroid impact at the time of the Late Cretaceous (66 Myr BP) mass extinctions (Alvarez 1986). But recent studies provide evidence that the Deccan Traps flood basalts in India, and perhaps the North Atlantic flood basalts, were erupted at the same time (Courtillot & Cisowski 1987, Officer et al 1987). Episodes of flood-basalt volcanism with peak periods lasting up to a few million years have occurred from time to time in the Earth's history. During the last 250 Myr, there were at least nine major flood-basalt episodes, some involving eruptions in more than one geographic area. When the ages of these flood-basalt episodes are subjected to a formal time-series analysis, they reveal a possible periodicity of roughly 30 Myr (Rampino & Stothers

1986). This is similar to the recent finding of a 26–32 Myr periodicity in the ages of biological mass extinctions (Fischer & Arthur 1977, Raup & Sepkoski 1984) and in the ages of episodes of impact cratering (Rampino & Stothers 1984, Alvarez & Muller 1984). Within the errors of dating, the ages of some of the flood basalts agree very well with the estimated ages of the mass extinctions and impact craters.

What can we infer from this? It may be that some massive outpourings of basalt are triggered by extraterrestrial impacts (Öpik 1958, Urey 1973, Rampino 1987). Or flood basalts might be generated by a quasi-periodic cycle of hotspot activity related to internal mantle dynamics (Loper & Stacey 1983). More speculatively, mass extinctions could be the result both of the aftereffects of a large impact and of related flood-basalt or explosive volcanism. In this case, cometary winters could be succeeded by volcanic winters that prolong the conditions adverse for life.

CONCLUSIONS

As has been shown in a number of studies, some of the largest historic eruptions are associated with atmospheric perturbations that have had a considerable impact on climate and agriculture. Even the greatest of these historic eruptions, however, was small compared with the very large explosive and effusive eruptions that are well known from the geologic record. A simple scaling-up of the effects of historic eruptions suggests that the much larger eruptions could have brought about severe, short-term coolings or "volcanic winters" over considerable portions of the globe. A very large eruption in the near future might have drastic effects on crop yields and could create food-supply crises in many areas, especially those regions of marginal productivity. Eruptions like these constitute a very real "volcanic hazard" in terms of the number of people that would be affected. There is no question that such large eruptions will recur, the only uncertainty lies in where and when.

Could individual "supereruptions" also lead to greatly prolonged climate cooling? The residence time of volcanic aerosols in the stratosphere is only of the order of a few years; therefore, one must invoke some form of positive climatic feedback to extend the cooling. One possibility is that a few cool summers could lead to significantly increased snow and ice cover at high latitudes, such that the increased albedo would further cool the Earth (see, e.g., Bray 1976). Both the Toba and Roza eruptions occurred at times of relatively rapid climate cooling, but no firm causal connection has been established. Although other large volcanic eruptions are not known to coincide with coolings, certain climatic regimes may be more sensitive to volcanic perturbations. In addition, climatic changes themselves may be able to influence the incidence and severity of volcanism (Rampino et al 1979). However, such connections remain elusive.

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