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31 Records of explosive volcanic eruptions over the last 500 years

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31.1 Introduction

Although explosive volcanic eruptions have long been suspected of having important effects on climate, the actual chronology of explosive events and their magnitude is not well known. Eruptions often occur in remote locations where, even today, they may go unrecorded (e.g. Sedlacek *et al.*, 1981; Mroz *et al.*, 1983). Only since the development of lidar observing stations in the 1970s have routine measurements of atmospheric aerosol loads been made, enabling specific aerosol clouds to be linked to preceding volcanic events (Reiter and Jager, 1986). Furthermore, the chemical composition of volcanic emissions is rarely well-documented and yet this is thought to be an important factor in determining what the consequent climatic effects might be (Rampino and Self, 1982; 1984; Devine *et al.*, 1984).

In this chapter we do not examine the climatic effects of explosive eruptions, but focus on the various chronologies of eruptions that have been constructed. The literature on climatic effects is large and there is little doubt that some eruptions in the past have affected climate over very large regions (for further discussions, see *inter alia*, Budyko, 1969; Lamb, 1970; Spirina, 1971; Yamamoto *et al.*, 1975; Oliver, 1976; Taylor *et al.*, 1980; Self *et al.*, 1981; Kelly and Sear, 1984; Angell and Korshover, 1985; Sear *et al.*, 1987; Kondratyev, 1988; Schonwiese, 1988; Bradley, 1988; Mass and Portman, 1989). The magnitude of any climatic effect depends on the volume of material ejected and the ejection height, the prevailing (and ensuing) stratospheric circulation pattern, and the chemical composition of the gas and tephra, particularly the amount of sulphur dioxide emitted. It appears that (at least for the major eruptions of the last century) effects on large-scale temperature averages were undetectable (that is, indistinguishable from noise) after 2-3 years (Kelly and Sear, 1984; Sear *et al.*, 1987; Bradley, 1988). However, others claim that eruptions may have had a significant impact on longer-term (lower frequency) changes of temperature (e.g. Budyko, 1969; Bryson and Goodman, 1980); indeed, there is some evidence that glacier advances over the last few centuries are closely linked to the cumulative atmospheric aerosol loading from volcanic eruptions (Bray, 1974; Porter, 1981, 1986).

Several attempts have been made to reconstruct the history of explosive volcanic eruptions over the last few centuries and four principal chronologies have been published. These are: a Dust Veil Index (DVI) (Lamb, 1970); a Volcanic Explosivity Index (VEI) (Simkin *et al.*, 1981; Newhall and Self, 1982); records of electrolytic conductivity or excess sulfate in ice cores (Hammer, 1977; Hammer *et al.*, 1980; Legrand and Delmas, 1987) and estimates of atmospheric optical depth (Pollack *et al.*, 1976; Bryson and Goodman, 1980). In addition, studies of frost damage in trees, historical and archeological records, and lunar eclipse data supplement and provide additional insight into these chronologies (LaMarche and

Hirschboeck, 1984; Stothers and Rampino, 1983; Keen, 1983). Several attempts have been made to improve on these approaches (e.g. Hirschboeck, 1980; Robock, 1981; Schonwiese, 1988) but the basic chronology of important eruptions is generally the same in each list. Here we discuss the derivation of these indices and their limitations; it will be apparent that none are ideal for climatic purposes.

31.2 Historical and geological records: the Dust Veil Index

Almost all studies of the climatic effects of volcanic eruptions have relied on Lamb's Dust Veil Index (DVI) chronology (Lamb, 1970). This was the first comprehensive effort to assess the probable climatic impact of volcanic eruptions and to construct a chronology of explosive eruptions and their magnitudes back to A.D. 1500. Unfortunately, the DVI assigned to individual eruptions is quite subjective and very dependent on observations and effects in mid-latitudes. Lamb based the DVI on historical and geological estimates of eruption magnitude, or on estimates of the volume of material dispersed into the atmosphere, or on volcanic effects on direct radiation receipts and surface temperature. These estimates were derived using one of the following formulae:

$$\mathrm{DVI} = 0.97\mathrm{R.E.t}, \quad \mathit{or} = 52.5\mathrm{T.E.t}, \quad \mathit{or} = 4.4\mathrm{q.E.t}$$

where R is the greatest percentage depletion of direct radiation, as registered by monthly averages in mid-latitudes of the hemisphere concerned,

T is the estimated lowering of annual temperature (in °C) over the mid-latitude zone of the hemisphere affected "for the year most affected",

q is the estimated total volume (in km^3) of solid matter dispersed as dust in the atmosphere,

and t is the total time in months between the eruption and last observation of the dust veil, or its effect on monthly radiation, or temperature values in mid latitudes.

The value of E ranges from 0.3 to 1.0, depending on the maximum extent attained by a dust veil. It is determined by the latitude of the eruption; for low latitude eruptions (20°N-20°S) E=1, for eruptions in the sub-tropics (20-35°N) E=0.7, for lower mid-latitude eruptions (35-42°N) E=0.5 and for higher latitude eruptions (>42°N) E=0.3.

The different multipliers used in each formula were derived empirically from a consideration of R, T and q for the 1883 eruption of Krakatau, so that the final DVI derived by each method would equal 1000 for this event. Lamb's 'final' estimate of the DVI for an individual eruption was based on an average of as many of these estimates as could be assembled. Global values of the DVI are given, as well as for each hemisphere separately, partitioning the dust between the hemispheres according to latitude of the eruption (see caption to Figure 31.1).

Values of R, T and t all depend on observations in mid-latitudes, though the effects of eruptions at different latitudes may be confined to the zone nearest to the eruption (Bradley, 1988). Furthermore, the derivation of the DVI using T (and an estimate of t based on temperature lowering) may lead to circular reasoning in any climatic analysis of the DVI. Lamb was cogniscent of this problem and made it clear which DVI values were derived wholly or partly in this way.

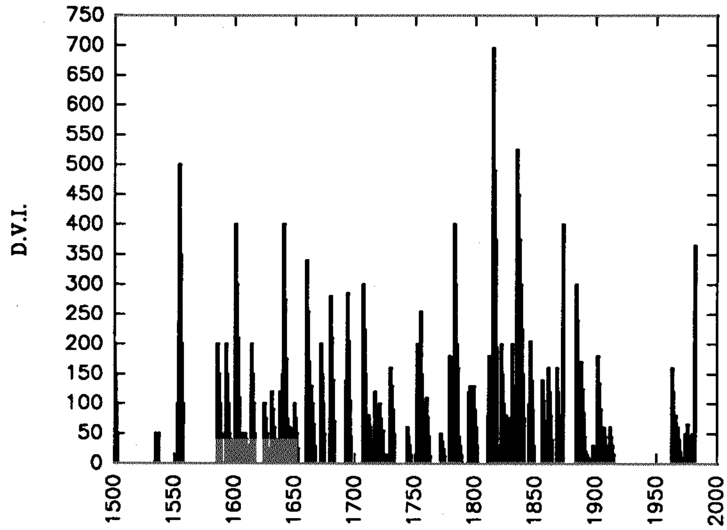

Figure 31.1 Cumulative DVI for the northern hemisphere, assuming the dust from an individual eruption is apportioned over four years with 40% of each DVI assigned to year 1, 30% to year 2, 20% to year 3 and 10% to year 4. Thus, the 1883 eruption of Krakatau (DVI=1000) results in values of 400 in 1883 declining to 100 in 1886. It is further assumed that all dust from eruptions poleward of 20°N remained in the northern hemisphere. For eruptions equatorward of 15°, the dust was assigned equally between the two hemispheres and for eruptions between 15 and 20°N and S, it was assumed that two thirds of the material remained in the hemisphere of the eruption and one third was dispersed to the other hemisphere (DVI values from Lamb, 1970, 1977, 1983).

When Lamb was unable to estimate a DVI based directly on any of the three formulae he assigned a DVI equal to the average for eruptions of similar size, using geological estimates of explosive eruption magnitude (largely based on Sapper, 1927). These were then adjusted (by E) for the latitude of the eruption. The averages were based on explosive eruptions which occurred after 1750 for which 'independent' DVI estimates (using the formulae above) could be made. Almost all of the DVI values before 1750 (and the majority of values from 1750 to 1900) were derived in this manner. When everything else failed, Lamb used historical records of unusual atmospheric phenomena ('volcanic sunsets'), plus his experience and knowledge of individual eruptions to produce a 'best estimate' of the eruption DVI.

In summary, Lamb's D.V.I. chronology is based on a number of different criteria, depending on the information available for a particular event. It is only partly objective and is biased towards the effects of eruptions on mid-latitudes. Nevertheless, Lamb's chronology is extremely useful if the methods used in its construction are clearly understood (cf. Kelly and Sear, 1982).

Figure 31.1 shows a time series of Lamb's cumulative DVI for the northern hemisphere, assuming the dust from an individual eruption is apportioned over four years, to simulate the gradual fall-out of dust from the atmosphere. The record is thus an estimate of the overall yearly volcanic aerosol loading of the atmosphere. It appears that dust loading was above

20th century levels for most of the preceding four centuries; highest levels prevailed in the early 19th century (1810s to 1830s) and in the 1780s, 1660s, 1630s and 1640s and in the 1600s. It is, of course, increasingly likely that as one goes back in time the cumulative values are only minimum estimates since many eruptions may not have been recognised.

Several attempts have been made to refine or improve upon Lamb's DVI. For example, Mitchell (1970) made a few minor changes to Lamb's estimates for the period 1850-1970 and Robock (1981) re-assessed the chronology from 1600 onwards. Robock eliminated as far as possible DVI values based partly or solely on temperature estimates and applied the time-distance decay model of Cadle *et al.*, 1976 (based on observed dust veil dispersal after the eruptions of Agung and Fuego) to produce a revised assessment of volcanic dust loading in time and space (Figure 31.2). The principal difference between Robock's estimates and

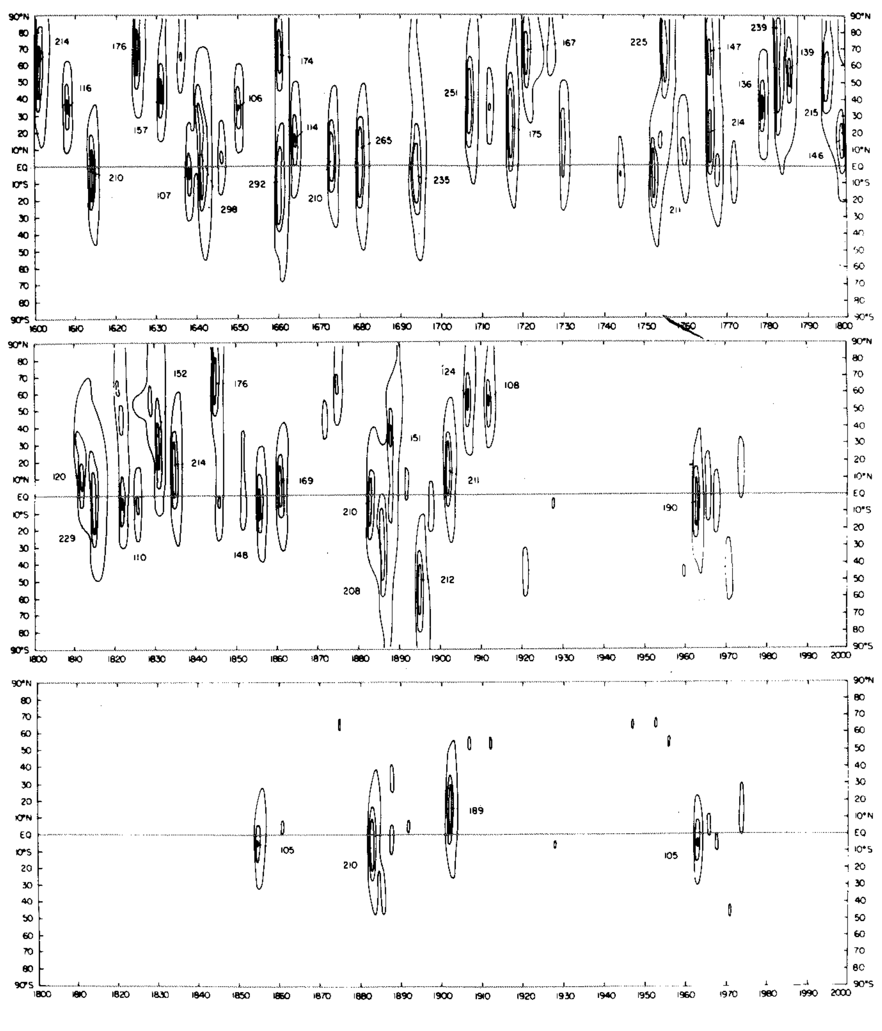

Figure 31.2 Latitudinal distribution of DVI as revised by Robock (1981). Contours are at index values of 20, 60 and 100 with the maximum value at the center also plotted (Robock, 1981).

Lamb's involve significantly lower DVI values for Krakatau (1883) Coseguina (1835) and Tambora (1815) and higher values for several events in the 17th and 18th centuries (Figure 31.3). However in all cases, the relative absence of major eruptions in the period after ~1920 is quite apparent.

31.3 Geological records: the Volcanic Explosivity Index

An alternative method of classifying explosive volcanic eruptions is that proposed by Newhall and Self (1982). They rank eruptions using only volcanological criteria to assess the magnitude, intensity, dispersive power and destructiveness of an event. The index does not rely on any assessment of temperature depression, atmospheric effects or reduction in radiation receipts, and is not weighted by climatic observations in mid-latitudes (as with the DVI or optical depth record) or high latitudes (as with the ice core record). It is thus a climatically independent estimate of explosivity, based primarily on geological criteria.

Each eruption is ranked from 1 to 8 (8 being the largest) and a chronology of volcanic events spanning the last 8,000 years has been constructed (Simkin *et al.*, 1981). The classification clearly identifies those events which are thought to have injected material into the stratosphere; such eruptions are assigned a Volcanic Explosivity Index (VEI) of 4 or more. These eruptions (described as Plinian or ultra-Plinian eruptions) produced at least 10^8 m^3 of ejecta and had column heights of 10-25km or more. Over the past 500 years there have been more than 110 eruptions with a VEI of 4 or more. These are listed in Table 31.1, divided into approximately equal area latitude bands in each hemisphere. Most major eruptions have occurred at high latitudes (>50°N) or within 10° of the Equator. There are relatively few records of large explosive eruptions in the southern hemisphere beyond the equatorial zone.

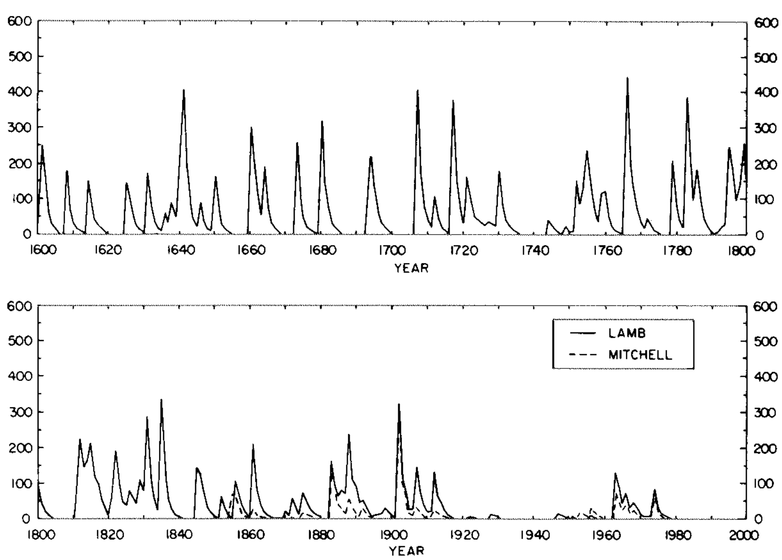

Figure 31.3 Annual average northern hemisphere volcanic dust veil indices as re-assessed by Robock (1981).

Table 31.1 Major Explosive Eruptions with VEI of 4 or more: 1500-1981 (after Simkin *et al.*, 1981).

Location	Elev. (m)	Lat.	Long.	Starting Date	VEI
A) Major High Latitude Eruptions (> 45 °N)					
Alaid, Kurile Is.	2339	50.8	155.5E	04 1981	4
Gareloi, Aleutians	1573	51.8	178.8W	08.1980	4
Mt.St.Helens, U.S.A.	2549	46.2	122.2W	05 1980	5
Bezymianny,Kamchatka	2800	56.1	160.7E	02 1979	4
Augustine, Alaska	1227	59.4	153.4W	01 1976	4
Plosky Tolbachik	3085	55.9	160.5E	07 1975	4
Sheveluch, Kamchatka	3395	56.8	161.6E	11 1964	4
Bezymianny,Kamchatka	2800	56.1	160.7E	03 1956	5
Spurr, Alaska	3374	61.3	152.3W	07 1953	4
Hekla, Iceland	1491	61.0	17.7W	03 1947	4
Sarychev, Kurile Is.	1497	48.1	153.2E	11 1946	4
Kliuchevskoi,Kamchatka	4850	56.2	160.8E	01 1945	4
Kliuchevskoi, Kamchatka	4850	56.2	160.8E	03 1931	4
Raikoke, Kurile Is.	551	48.3	153.3E	02 1924	4
Katla, Iceland	1363	63.6	19.0W	10 1918	4
Katmai, Alaska	841	58.3	155.2W	06 1912	6
Ksudach, Kamchatka	1079	51.8	157.5E	03 1907	5
Thordarhyna, Iceland	1659	64.3	17.6W	05 1903	4
Augustine, Alaska	127	59.4	153.4W	10 1883	4
Askja, Iceland	1510	65.0	16.8W	03 1875	5
Grimsvotn, Iceland	1719	64.4	17.3W	01 1873	4
Sinarka, Kurile Is.	934	48.9	154.2E	? 1872	4
Sheveluch, Kamchatka	3395	56.8	161.6E	02 1854	5
Chikurachki, Kurile Is.	1817	50.3	155.5E	12 1853	4
Hekla, Iceland	1491	64.0	19.7W	09 1845	4
Isanotski, Aleutian Is.	2446	54.8	163.7W	03 1825	4
Beerenberg, Jan Mayen	2277	71.1	8.2W	? 1818	4
St. Helens, U.S.A.	2549	46.2	122.2W	? 1800(D)	4
Pogromni, Aleutians Is.	2002	54.6	164.7W	? 1795	4
Alaid, Kurile Is.	2339	50.8	155.5W	* 1793	4+
Laki, Iceland	500	64.1	18.3W	* 1784	4
Raikoke, Kurile Is.	551	48.3	153.3E	? 1778	4
Hekla, Iceland	1491	64.0	19.7W	04 1766	4
Katla, Iceland	1363	63.6	19.0W	10 1755	5
Oraefajokull, Iceland	2119	64.0	16.7W	08 1727	4
Katla, Iceland	1363	63.6	19.0W	05 1721	4
Chirpoi, Kurile Is.	624	46.5	150.9E	12 1712	4?
Hekla, Iceland	1491	64.0	19.7W	02 1693	4
Chikurachki, Kurile Is.	1817	50.3	155.5E	? 1690(T)	4
Hekla, Iceland	1491	64.0	19.7W	07 1510	4
St. Helens, U.S.A.	2549	46.2	122.2W	? 1500(D)	5
B) Major Mid Latitude Eruptions (20-45 °N)					
Tiatia, Kurile Is.	1822	44.4	146.3E	07 1973	4
Komaga-take, Japan	1140	42.1	140.7E	06 1929	4
Sakura-jima, Japan	1118	31.6	130.7E	01 1914	4
Tarumai, Japan	1024	42.7	141.4E	03 1909	4
Suwanose-jima, Japan	799	29.5	129.7E	10 1889	4
Bandai, Japan	1819	37.6	140.1E	07 1888	4
Nasu, Japan	1917	37.1	140.0E	07 1881	4
Suwanose-jima, Japan	799	29.5	129.7E	? 1877	4

Table 31.1 – *continued*

Location	Elev. (m)	Lat.	Long.	Starting Date	VEI
Komaga-take, Japan	1140	42.1	140.7E	09 1856	4
Usu, Japan	725	42.5	140.8E	04 1853	4
Usu, Japan	725	42.5	140.8E	03 1822	4
Asama, Japan	2550	36.4	138.5E	05 1783	4
Komage-Take, Japan	1140	42.1	140.7E	? 1765	4
Oshima-O-Shima, Japan	714	41.5	139.4E	08 1741	4
Tarumai, Japan	1024	42.7	141.4E	08 1739	5
Fuji, Japan	3776	35.4	138.7E	12 1707	4
Iwate, Japan	2041	39.9	141.0E	02 1686	4
Tarumai, Japan	1024	42.7	141.4E	08 1667	5
Usu, Japan	725	42.5	140.8E	08 1663	5
Agua de Pau, Azore Is.	948	37.8	25.5W	06 1563	4

C) Major Low Latitude Eruptions, (0° to 20°N)

Mt.Pagan,Mariana Is.	570	18.3	145.8E	05 1981	4
Fuego, Guatemala	3763	14.5	90.9W	10 1974	4
Awu, Indonesia	1320	3.7	125.5E	08 1966	4
Taal, Philippines	400	14.0	121.0E	09 1965	4
Fuego, Guatemala	3763	14.5	90.9W	01 1932	4
Agrigan, Mariana Is.	965	18.8	145.7E	04 1917	4
Colima, Mexico	4100	19.4	103.7W	01 1913	4?
Taal, Philippines	400	14.0	121.0E	01 1911	4
Santa Maria,Guatemala	3772	14.8	91.6W	10 1902	6
Soufriere,West Indies	1178	13.3	61.2W	05 1902	4
Pelee, West Indies	1397	14.8	61.2W	05 1902	4
Dona Juana,Colombia	4250	1.5	76.9W	11 1899	4
Purace, Columbia	4600	2.4	76.4W	10 1869	4
Purace, Colombia	4600	2.4	76.4W	12 1849	4
Coseguina, Nicaragua	859	13.0	87.6W	01 1835	5
Colima, Mexico	4100	19.4	103.7W	02 1818	4
Mayon, Philippines	2462	13.3	123.7E	02 1814	4
Soufriere, St. Vincent	1178	13.3	61.2W	04 1812	4
San Martin, Mexico	1550	18.6	95.2W	03 1793	4
Jorullo, Mexico	1330	19.0	101.7W	* 1764	4
Tongkoko, Indonesia	1149	1.5	125.2E	? 1680	4
Gamkanora, Indonesia	1635	1.4	127.5E	05 1673	4
San Salvador, El Salvador	1850	13.7	89.3W	? 1671	4
San Salvador, El Salvador	1850	13.7	89.3W	? 1575	4?
Arenal, Costa Rica	1552	10.5	84.7W	? 1525	4

D) Major Southern Hemisphere Eruptions: Low Latitudes (0-20°S)

Ulawun, New Britain	2300	5.0	151.3E	10 1980	4
Negra, Galapagos Is.	1490	0.8	91.2W	11 1979	4
Fernandina,Galapagos	1495	0.4	91.6W	06 1968	4
Lengai, E. Africa	2886	2.8	35.9E	08 1966	4
Kelut, Indonesia	1731	7.9	112.3E	04 1966	4
Agung, Indonesia	3142	8.3	115.5E	03 1963	4
Bagana, Solomon Is.	1702	6.1	155.2E	02 1952	4
Ambryn, New Hebrides	1334	16.3	168.1E	* 1951	4
Lamington, New Guinea	1780	8.9	148.2E	01 1951	4

Table 31.1 – *continued*

Location	Elev. (m)	Lat.	Long.	Starting Date	VEI
Rabaul, New Britain	229	4.3	152.2E	05 1937	4
Manam, New Guinea	1725	4.1	145.1E	08 1919	4
Tungurahua, Ecuador	5016	1.5	78.5W	04 1918	4
Krakatau, Indonesia	813	6.1	105.4E	08 1883	6
Galunggung, Indonesia	2168	7.4	108.1E	10 1822	5
Tambora, Indonesia	2851	8.3	118.0E	04 1815	7
Papandayan, Indonesia	2665	7.4	107.7E	08 1772	4
Cotopaxi, Ecuador	5897	0.8	78.4W	04 1768	4
Cotopaxi, Ecuador	5897	0.8	78.4W	11 1744	4
Long Is., New Guinea	1304	5.4	147.1E	? 1700(C)	6
Quilotoa, Ecuador	3914	0.9	78.9W	11 1600	4?
Kelut, Indonesia	1731	7.9	112.3E	? 1586	4
Cotopaxi, Ecuador	5897	0.8	78.4W	06 1534	4
E) Major Southern Hemisphere Eruptions: Mid and High Latitudes (>20°S)					
Nilahue, Chile	400	40.4	72.1W	07 1955	4
Azul, Chile	3810	35.7	70.8W	04 1932	6
Puyehue, Chile	2240	40.6	72.1W	12 1921	4
Tarawera,New Zealand	1111	38.2	176.5E	06 1886	5
Peteroa, Chile	4090	35.3	70.6W	12 1762	4

* Continuous eruptions over one or more years.
A letter in parenthesis or ? after date indicates date uncertain; eruption dated dendrochronologically (D) or tephrochronologically (T) or by radiocarbon (C).

However, there is no doubt that this catalog omits or underestimates the size of many climatically significant eruptions which must have occurred in remote areas such as the Aleutians, and Kamchatka, or in the Andes, New Zealand and Antarctica. Those volcanoes which have been studied most carefully (e.g. Hekla in Iceland and Taupo in New Zealand) account for a large fraction of the major eruptions which are known to have occurred (Simkin *et al.*, 1981) and, no doubt, as further geological studies are carried out the list of explosive eruptions will expand, particularly for the period before about 1850.

Of particular relevance to the climatic effects of explosive eruptions is the total volume of sulfur-rich volatiles (such as SO_2 and H_2S) which is emitted (Rampino and Self, 1984). These gases result in sulphuric acid aerosols being produced in the stratosphere and such aerosols are known to be important in reducing solar radiation receipts at the surface (Toon and Pollack, 1984). Various estimates have been made of the H_2SO_4 'yield' from major eruptions; Table 31.2 lists recent results based on petrologic studies of glass inclusions in tephra from a number of the largest eruptions of the last 500 years, together with their corresponding VEI (Devine *et al.*, 1984; Rampino and Self, 1984; Palais and Sigurdsson, 1989). This provides a quite different perpective on the climatic significance of these eruptions. In terms of H_2SO_4 yield, the most important eruption of the last 500 years was the fissure eruption of Laki in Iceland (1783). This is the well-known event which produced 'dry fogs' over Europe, as described by Franklin (1784) and discussed at length by Lamb (1970). The Laki eruption

Table 31.2 Petrologic estimates of volatile emissions based on glass inclusion analysis (data from Rampino and Self, 1984; Symonds *et al.*, 1988 and Palais and Sigurdsson, 1989).

Eruption	**Date**	**Lat.**	**VEI**	**H_2SO_4 (metric tons)**
St. Helens, U.S.A.	1530	46°N	5	2.30×10^5
Laki, Iceland	1783	64°N	4	9.03×10^7
St. Helens, U.S.A.	1800	46°N	4	3.50×10^3
Tambora, Indonesia	1815	8°S	7	5.24×10^7
Coseguina,Nicaragua	1835	13°N	5	0
Krakatau, Indonesia	1883	6°S	6	2.94×10^6
Tarawera, New Zealand	1886	38°S	5	5.00×10^6
Santa Maria, Guatemala	1902	15°N	6	1.80×10^5
Soufriere, St. Vincent	1902	13°N	4	2.40×10^5
Katmai, Alaska	1912	58°N	6	7.90×10^6
Bezymianny, Kamchatka	1956	56°N	5	6.00×10^6
Agung, Indonesia	1963	8°S	4	2.84×10^6
St. Helens, U.S.A.	1980	46°N	5	7.90×10^4
El Chichon, Mexico	1982	17°N	4	7.00×10^4

(VEI=4) was comparable in H_2SO_4 production to Tambora which has a VEI rating of 7. Tambora is considered to have been the most violent explosive eruption of the Holocene; about 50km^3 of magma erupted within 24 hours, some of which probably reached the upper stratosphere (Self and Rampino, 1984; Stothers, 1984). By contrast, Laki was a very extensive (>560km^2) non-explosive fissure eruption which continued for about 8 months. It is possible that intense convection cells associated with an eruption of this size could have led to stratospheric injection of gases and aerosols leading to similar climatic consequences as a more explosive eruption (Wolff *et al.*, 1984; Stothers *et al.*, 1986). In contrast to Tambora and Laki, Coseguina tephra reveals little evidence of sulfur-rich volatile emissions, though very large quantities of HCl were produced (1×10^7 metric tons) (Palais and Sigurdsson, 1989). The implications of such cloride-rich eruptions for climate are as yet unclear, but it seems likely that they have a major influence on stratospheric ozone concentrations (Johnson, 1980; Symonds *et al.*, 1988)

Table 31.2 also shows that the eruptions of Tarawera (1886) Katmai (1912) Bezymianny (1956) and Agung (1963) all produced more H_2SO_4 than Krakatau (1883) which explains why these events are so clearly noticeable in continental temperature records (Bradley, 1988). Indeed, there is a power relationship between petrologic estimates of sulfur yield and estimated temperature decrease over the northern hemisphere following major explosive eruptions over the last few centuries (Palais and Sigurdsson, 1989) (Figure 31.4).

31.4 Glaciological records

Large variations in the electrolytic conductivity of Greenland ice cores were first shown to be related to the deposition of acidic snow following large explosive eruptions, by Hammer (1977) and Hammer *et al.* (1980). Volcanic eruptions may produce large quantities of sulfur

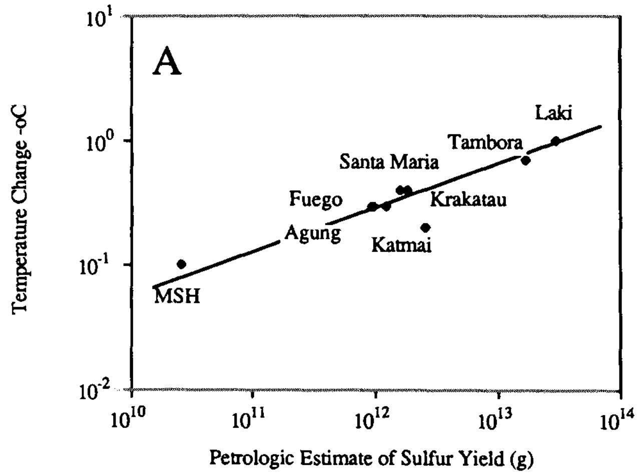

Figure 31.4 Relationship between petrologic estimates of sulfur yield (in grams) and estimates of northern hemisphere continental temperature decreases (°C) following the explosive eruptions indicated (Palais and Sigurdsson, 1989).

and chlorine gases which are converted to acids in the atmosphere. The resulting acidic snowfall (and dry deposition of acidic particles directly on the ice sheets) produces high levels of conductivity or 'spikes' above natural background levels. In most cases these acidity spikes result from excess sulphuric acid events (Figure 31.5). Hammer's original studies showed a remarkable similarity between electrolytic conductivity in the Greenland Crete ice core and Lamb's DVI, indicating that the elevated acidity (above background levels) could be used as an index of volcanic explosivity. This concept has now been extensively investigated in other cores from Greenland, Antarctica and elsewhere, with careful analyses to determine the precise chemistry of the acidity spikes (e.g. Holdsworth and Peake, 1984; Mayewski *et al.*, 1986; Legrand and Delmas, 1987; Lyons *et al.*, 1990). This enables more precise 'fingerprinting' of individual eruptions, some of which produce large amounts of HCl, for example, rather than H_2SO_4 (see discussion above).

Recently, Legrand and Delmas (1987) proposed that a Glaciological Volcanic Index (GVI) be compiled from ice core measurements in different parts of the world. Several problems face such a development. Firstly, the ice core records are primarily from high latitudes and are thus strongly biased towards high latitude eruptions, particularly (in the case of Greenland ice cores) those of Icelandic volcanoes (Hammer, 1984). Secondly, the eruption signal may differ significantly in magnitude from one ice core to another and important events may not appear at all in some ice core records (Delmas *et al.*, 1985). This reflects the fact that the spatial distribution of acidic snowfall varies after volcanic eruptions and hence the record of deposition will vary from one ice coring site to another (Clausen and Hammer, 1988). Thirdly, in some low latitude ice cores, deposition of alkaline aerosols may neutralise the volcanic acids and hence eliminate the eruption signal. These problems do not seem to be insuperable; more ice cores from lower latitudes (cf. Thompson, this volume) will help in constructing a complete catalog of explosive eruptions and in resolving latitudinal effects on

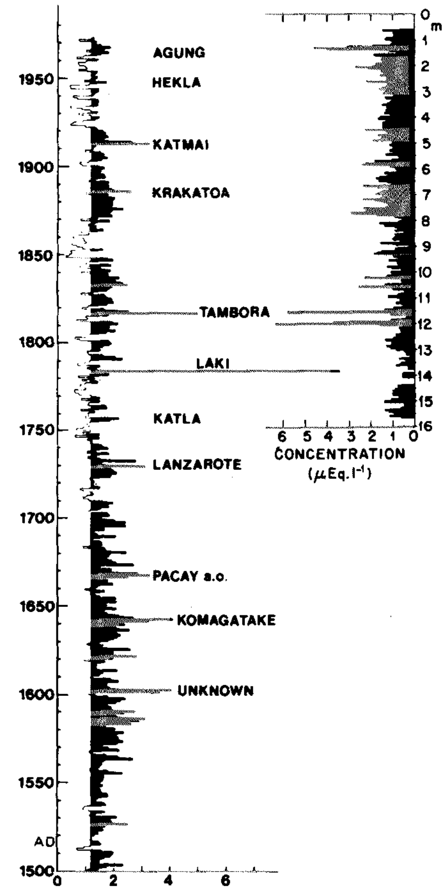

Figure 31.5 Mean acidity of annual layers from A.D. 1500 to 1972 in the ice from Crete, Greenland (left) and the excess sulfate record from Dome C, Antarctica for the last ~200 years (right). Acidity values in the Greenland core in excess of 1.2 equiv H^+ per kg ice are considered to be due to the fallout of volcanic acids (Greenland record from Hammer *et al.*, 1980; Dome C record from Legrand and Delmas, 1987).

the dispersal of volcanic material. The collection of many short cores from a wide area of the larger ice sheets, and along polar/alpine transects (e.g. from the South Pole to Peru) will enable a better assessment of the spatial pattern of acid deposition to be made (e.g. Clausen and Hammer, 1988; Mulvaney and Peel, 1987).

More detailed chemical studies thus hold the promise of a significant improvement in our understanding of volcanic explosivity through time. However, at present there is no comprehensive GVI catalog for eruptions, and extreme values in the published ice core conductivity profiles do not differ substantially from extremes in the DVI and VEI for the last 500 years.

31.5 Tree ring records

Major explosive eruptions may bring about changes in the atmospheric circulation that are sufficiently anomalous to seriously disrupt normal growing conditions for plants. In the case of trees, such anomalies may result in a sequence of extremely thin growth layers, or (in the most extreme cases) actual tissue damage to the growing cells.

In a study of frost ring damage to trees in the western United States, LaMarche and

Hirschboeck (1984) identified 17 times in the last 500 years when frost damage occurred in two or more regions of the western U.S. These are listed in Table 31.3. Many of these correspond to (or immediately follow) years in which major explosive eruptions occurred, according to Lamb (1970). Less correspondence is observed with the VEI chronology of Simkin *et al.* (1981). Table 31.2 also lists years with extremely narrow rings, or ring width sequences, in temperature-sensitive Pinus balfouriana from the Sierra Nevada of California (Scuderi, 1990). These years are assumed to be related to anomalous climatic conditions following major eruptions. Some correspondence with the frost-damaged tree ring record is apparent (e.g. 1601, 1640) but additional years of severe growth reduction are also identified.

When widespread conditions of frost-damage or narrow growth increments correspond to other independent records of explosive eruptions, the tree rings provide additional proxy evidence for the eruption and indicate the importance or magnitude of the explosive events

Table 31.3 Years with notable frost ring damage to trees in the western United States (1500-1970)* (LaMarche and Hirschboeck, 1984) and years with extremely narrow rings or ring width sequences in Sierra Nevada *Pinus balfouriana* (Scuderi, 1990).

Frost rings	Narrow rings
1965	
1941	
	1913
1912	
1902	
	1884
1866	
1837	
1831	
1828	
1817	
	1815
1805	
	1784
1761	
1732	
	1730
	1725
	1666
1660	
1640	1640
1601	1601

*Notable frost rings are those occurring in 50% or more of sampled trees in any one location and at two or more sites. Sites range from eastern California to New Mexico and Colorado.

through their impact on the general circulation. This information can be useful in refining the chronology of explosive eruptions and perhaps point to significant gaps in other volcanic chronologies. However, it is clear that not all growth damage or extremely narrow growth increments can be expected to result from volcanic eruptions; conversely, not all eruptions will produce reduced growth or tissue damage. Indeed, in some areas, circulation anomalies may result in *enhanced* growth (*cf.* Lough and Fritts, 1987). Hence, the records from trees must be considered as of only limited use in constructing a comprehensive chronology of explosive volcanic eruptions. However, assembling records of frost-damaged trees from different regions, together with studies of extremely narrow tree ring sequences (e.g. Baillie and Munro, 1988) would make it possible to determine which volcanic events were of greatest ecological significance. Attention can then be focused on the chemistry and dynamics of these events, and of the subsequent changes in atmospheric circulation which they brought about.

31.6 Instrumental records

Ideally, the optimum index of volcanically-induced turbidity would be a time series of direct radiation receipts at a set of well-distributed, high altitude sites. Unfortunately, such a data set does not exist. There have been actinometric (solar radiation) measurements in various locations around the world since the 1880s, but changes in instrumentation and the lack of fixed, long-term observations limit the usefulness of the records. Nevertheless, the available data provide an interesting comparison with the DVI, VEI and glaciological records discussed above. Figure 31.6 shows a composite record of solar radiation receipts at 20 observing stations in the northern hemisphere (between 32° and 62°N) (Asaturov *et al.*, 1986). This is similar to the optical depth record derived by Bryson and Goodman (1980) from 42 actinometric and pyrheliometric records between 20° and 65°N (Figure 31.6). Optical depth is a measure of the size and number of particles in a column of air (Toon and Pollack, 1980). Both compilations show sharp decreases in solar radiation receipts following major explosive eruptions such as Krakatau (1883) Santa Maria (1902) and Katmai (1912). Interestingly, both records also indicate a decrease in solar radiation (increase in optical depth) since ~1940, accentuated by the eruptions of Agung (1963) and El Chichon (1982). Soviet analysts (e.g. Budyko, 1969; Pivovarova, 1977) have argued for many years that the post-1940s cooling of the northern hemisphere parallels this decline in direct radiation. However, the reasons for the gradual increase in optical depth since 1940 are not clear; no comparable increase in volcanic activity is indicated by the DVI or VEI catalogs, though there is a small increase in Greenland ice core acidity over this period. Changes in tropospheric turbidity due to anthropogenic activities may be a factor in the optical depth increase.

Further research is needed to determine if these composite records accurately indicate volcanic aerosol loading over the last century. However, there is no propect of extending the record further back in time.

31.7 Conclusions

The record of large explosive eruptions since A.D. 1500 is probably quite incomplete, making it difficult to assess their overall impact on climate There is no single index of volcanic

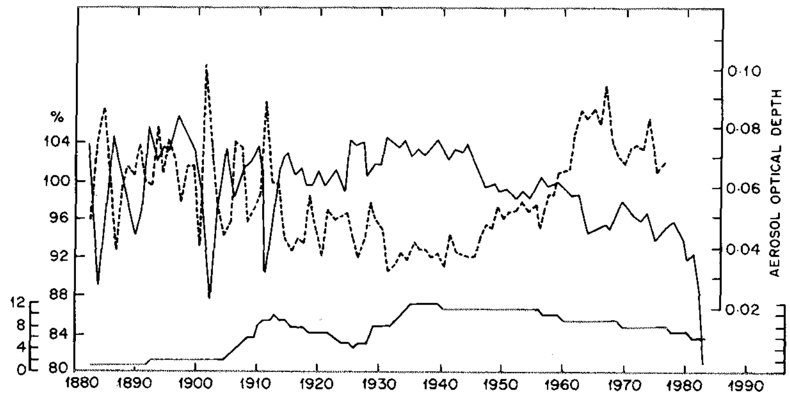

Figure 31.6 *Upper solid line*: variations in annual solar radiation receipts 1883-1983 at 20 actinometric radiation stations between 32° and 62°N as a percentage of the long-term mean for all records. *Lower solid line* shows the number of stations operating in each year; before 1905 only 2 stations were available (Pavlovsk/Leningrad and Montpelier). Similarly, after 1970, the average is based on 7 Soviet stations and Madison, Wisconsin (after Asaturov *et al.*, 1986).

Dashed line: mean annual residual aerosol optical depth, 1883-1975, based on a network of 42 stations between 20° and 65°N. This is the optical depth obtained after adjusting the data for the effects of clean air, water vapor, ozone etc., which amounts to ~0.212 in the northern hemisphere (Bryson and Goodman, 1980).

explosivity which is ideal for climatic studies since every index has its own limitations. In all catalogs, it is likely that some eruptions have been missed and the magnitudes of others have been mis-classified. However, there is little disagreement between the various catalogs about which were the biggest eruptions of the last 500 years, though geochemical studies reveal important differences in the potential climatic significance of explosive eruptions; not all large eruptions had the same potential for climatic effects. Only by combining geological studies with glaciological and other lines of evidence can a more complete assessment of volcanic explosivity through time be obtained.

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