Correlations between eruption magnitude, SO2 yield, and surface cooling

S. Blake

Geological Society, London, Special Publications 2003; v. 213; p. 371-380
doi:10.1144/GSL.SP.2003.213.01.22

Correlations between eruption magnitude, SO_2 yield, and surface cooling

S. BLAKE

Department of Earth Sciences, The Open University, Walton Hall, Milton Keynes, MK7 6AA, UK.
(e-mail: s.blake@open.ac.uk)

Abstract: Sulphurous gases from explosive eruptions have the potential to form stratospheric aerosols and so produce surface cooling on a hemispheric to global scale. However, testing for any correlation between SO_2 yield and surface cooling is hampered by instrumental SO_2 and temperature measurements being available for time periods that include only a few large eruptions. To overcome this, published dendroclimatological data, satellite (Total Ozone Mapping Spectrometer) data on SO_2 emissions, stratospheric optical depth data, and volcanological observations are integrated, revealing several relevant new correlations. First, the efficient conversion of SO_2 into stratospheric aerosols occurs when the ratio of plume height to tropopause height is greater than about 1.5. Second, the mass of emitted SO_2 correlates well with the mass of erupted magma. The SO_2 yield is 0.1 to 1% by mass of magma, irrespective of composition. The best-fit power law (r^2=0.67) is mass of SO_2 in Mt=1.77(mass of magma in Gt)$^{0.64}$. Third, of the eruption clouds that are believed to have entered the stratosphere in the period 1400–1994, those with masses <5 Gt magma (DRE <2 km^3) appear to have had insignificant effects on Northern Hemisphere summer temperature. The scattered data for eruptions of >10 Gt (>4 km^3) magma suggest a mean cooling effect of about 0.35 °C.

Volcanic gases can influence the Earth's surface environment in many ways. One of these is the surface cooling in the one or two years following large volcanic eruptions. Notable examples of this effect are the unusually cold weather conditions reported after the 1815 eruption of Tambora (Stommel & Stommel 1983; Stothers 1984) and the instrumental record of a decreased global mean surface temperature after the 1991 eruption of Pinatubo (McCormick *et al.* 1995; Self *et al.* 1996). The cooling mechanism has been described lucidly by Peter Francis (Francis 1993; Colling *et al.* 1997) and reviewed most recently by Robock (2000). For sufficiently high eruption rates, tephra and gas are lofted into the stratosphere, where volcanic SO_2 gas is converted to H_2SO_4 aerosol droplets and dispersed. The aerosol layer increases the optical depth of the stratosphere and reflects some solar radiation, leading to cooling of the troposphere. The increase in optical depth is predicted to increase with the mass of SO_2 (Stothers 1984; Pinto *et al.* 1989) and the radiative forcing is predicted to be proportional to the increase in optical depth (Pollack *et al.* 1976; Lacis *et al.* 1992; Andronova *et al.* 1999; Grieser & Schönwiese, 1999). Therefore, for those eruptions that penetrate the stratosphere, surface cooling should correlate with the mass of SO_2 released. Understanding this correlation is important for estimating the likely effects of poorly documented historic eruptions, assessing the environmental impacts of extremely large prehistoric eruptions, and for predicting climatic and therefore agricultural responses to any future eruptions that are larger than those experienced in historic times. The purpose of this chapter is to investigate this proposed correlation, using data from historical eruptions.

Devine *et al.* (1984), Rampino and Self (1984) and Palais and Sigurdsson (1989) gave evidence that sulphur-rich eruptions caused decreases in global mean surface temperature of 0.1 to 0.5 °C. However, they estimated the SO_2 by using a petrological method that, as will be discussed below, is now known to underestimate the sulphur yield by up to two orders of magnitude in some cases (e.g. Westrich & Gerlach 1992). In addition, instrumental data on global and hemispheric mean surface temperature are reliable only after the 1850s, so that important eruptions such as Tambora 1815 pre-date the period of good temperature data. The correlation between SO_2 and cooling proposed by Devine *et al.* (1984) and Palais and Sigurdsson (1989) was partly based on eruptions that apparently caused cooling of about 0.1 °C, but this is similar to the uncertainty in mean surface

From: OPPENHEIMER, C., PYLE, D.M. & BARCLAY, J. (eds) *Volcanic Degassing*. Geological Society, London, Special Publications, **213**, 371–380. 0305-8719/03/$15.00

temperature estimates (Hansen & Lebedeff 1987). It is therefore appropriate to re-evaluate the link between volcanic SO_2 emissions and surface cooling.

Data selection

Seeking to define the nature of any correlation between SO_2 yield and surface cooling presents a number of challenges. These include the estimation of the mass of SO_2 gas released by an eruption and the change (if any) in mean surface temperature following an eruption. Allied to the latter point is the question of isolating a volcanic signal from all the other random and non-random controls on temperature (e.g. Kelly & Sear 1984; Mass & Portmann 1989; Robock & Mao 1995; Sadler & Grattan 1999). Furthermore, temperature and SO_2 data must be available for as many eruptions as possible.

Briffa *et al.* (1998) showed that the density of late growth tree rings is a good proxy for mean summer land temperature, and they presented a tree ring record of Northern Hemisphere summer temperature for the period 1400 to 1994. They give values of the difference between annual tree ring density and the mean 1881–1960 tree ring density, normalized by the standard deviation (s.d.) of the 1881–1960 data. 1 s.d. is equivalent to a temperature difference from the mean of 0.117 °C. Most of the tree ring data lies within ±2 s.d. of the mean; few values are greater than +2 standard deviations, but several are less than −2 s.d. (Fig. 1). Briffa *et al.* (1998) showed that many of the extreme negative temperature anomalies are associated with large volcanic eruptions (see also Jones *et al.* 1995; Pyle 1998). The NHD1 series of Briffa *et al.* (1998) (http://www.ngdc.noaa.gov/paleo.html) is therefore a useful data-set with which to further investigate volcano–climate linkages. One reason for this is that it covers a time-span that is much longer than the instrumental hemispheric or global mean surface temperature record (1860 onwards). This allows the effects of many eruptions to be investigated with a common climate indicator. In addition, because the potential climatic effects of volcanic eruptions may damage crop production, it is useful to measure climate impact using an index that is directly relevant to crop growth. The NHD1 record of mean summer surface land temperature satisfies this requirement more than, say, annual or global temperature data.

The mass of SO_2 released by volcanic eruptions is often estimated by a petrological method, based on the mass of magma erupted and the difference in S content of melt inclusions that were trapped in phenocrysts prior to eruption, and degassed matrix glass (Devine *et al.* 1984). However, independent estimates of the amount of SO_2 released in several recent eruptions are now available from Total Ozone Mapping Spectrometer (TOMS) instruments with detection limits ranging from 5 to 20 kt and an uncertainty of ±30% (Bluth *et al.* 1997; Schnetzler *et*

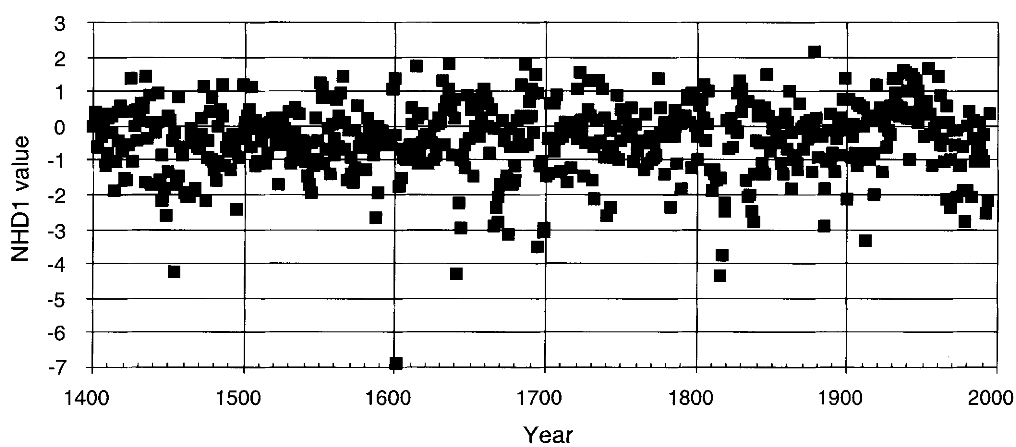

Fig. 1. The NHD1 data of Briffa *et al.* (1998) (http://www.ngdc.noaa.gov/paleo.html) showing Northern Hemisphere summer temperature anomalies, with respect to the 1881–1960 mean, in standard deviation units for the period 1400 to 1994. 1 s.d. unit represents 0.117 °C. Note the lack of extreme positive values. Extreme negative values coincide with known eruptions, in particular those of Kuwae 1452, Huaynaputina 1600, Mount Parker 1641, Tambora 1815, Krakatau 1883, and Novarupta 1912.

al. 1997; Krueger *et al.* 2000). In cases where the two methods can be compared, the petrological method has so far been found to underestimate the SO_2 yield substantially (Fig. 2). The reason for the discrepancy is probably related to the assumption in the petrologic method that syn-eruptive melt-degassing is the only source of erupted SO_2. The most likely source of the 'excess' sulphur is a S-bearing vapour in the pre-eruptive magma (Westrich & Gerlach 1992; Scaillet *et al.* 1998; Wallace 2001), with possible additional contributions from un-erupted magma (Andres *et al.* 1991) and hydrothermal sources (Oppenheimer 1996). So, although the petrological method is relatively easy to apply, its results give only a minimum estimate of SO_2 yield, and must be treated with caution. The most robust estimate of SO_2 yield is therefore thought to be TOMS measurements of SO_2 present in the atmosphere. Values have been compiled for eruptions in the period 1979 to 1994 by Bluth *et al.* (1993), Symonds *et al.* (1994) and Bluth *et al.* (1997). Data from TOMS and NHD1 are therefore used to underpin the investigation reported in this chapter.

Correlation between mass of SO_2 and optical depth of the stratosphere

Stothers (1984) derived a relationship between the mass of stratospheric aerosol, M_D, in grams, and the resultant increase in optical depth, $\Delta\tau$

$$\Delta\tau = 6.5 \times 10^{-15} M_D \quad (1)$$

The aerosol particles are assumed to have a chemical composition of 75% H_2SO_4 and 25% H_2O by mass (Toon & Pollack 1973), so equation 1 can be rewritten in terms of the mass of SO_2 responsible for the aerosols, M_{D,SO_2}:

$$\Delta\tau = 0.013\ M_{D,SO_2} \quad (2)$$

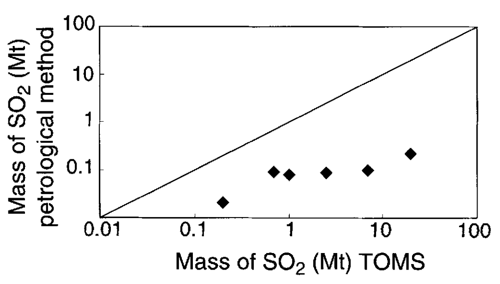

Fig. 2. Comparison of petrological and TOMS estimates of volcanic SO_2 yields. Data and data sources in Table 1.

where M_{D,SO_2} is expressed in megatonnes (1 Mt$=10^{12}$ g).

Of the total mass of SO_2 released by an eruption, M_{SO_2}, only some fraction will end up in the stratosphere and be available to form aerosol particles. Plumes that rise only into the lower troposphere should have a negligible effect on stratospheric aerosols ($M_{D,SO_2}/M_{SO_2} \approx 0$), whereas plumes that rise far into the stratosphere will transfer most of the SO_2, such that $M_{D,SO_2} \approx M_{SO_2}$. The ratio $\Delta\tau/M_{SO_2}$ should approach 0.013 as the ratio of plume height to tropopause height increases above 1. Data with which to test this are available for nine plumes and bear out the prediction (Table 1 & Fig. 3). Plumes that reach heights of more than about 1.5 times the tropopause height transport essentially all of their SO_2 to the stratosphere. Lower eruption clouds that still breach the tropopause will have a lesser effect on aerosol production (all else being equal) than predicted by equation 1.

Correlation between mass of SO_2 and mass of magma

As already discussed, the petrological method of estimating M_{SO_2} does not always give results that are consistent with TOMS data. Using the petrological method is unlikely, therefore, to provide adequate values of M_{SO_2} with which to compare the summer cooling signal in the 1400 to 1994 NHD1 record. An alternative means of estimating M_{SO_2} is required. In Figure 4, M_{SO_2} from TOMS data is plotted against the mass of magma erupted, and shows that the SO_2 released is usually between 0.1 and 1% by mass of the magma erupted (equivalent to 500 to 5000 ppm S). The good correlation is similar to those given by Oppenheimer (1996) and Wallace (2001), although these authors included a mixture of petrological, COSPEC, and selected TOMS estimates on their diagrams. Using exclusively TOMS data, Pyle *et al.* (1996) and Schnetzler *et al.* (1997) found good correlations between M_{SO_2} and volcanic explosivity index (VEI). VEI is a compound measure of an eruption's characteristics (Newhall and Self 1982), with the result that some eruptions may be difficult to classify uniquely. Carey and Sigurdsson (1989) found that several Plinian eruptions would be classified differently on the basis of column height or mass erupted. So, using VEI as a predictor of an individual eruption's SO_2 yield carries a potentially large uncertainty. Eruption mass (or volume) is a more precise measure of eruption size, so Figure 4 should present the best chance of assessing any correlation between eruption size and M_{SO_2}.

Table 1. *Data for selected eruptions detected by TOMS.*

Volcano	Date	Mass of magma (Gt)	Plume height above sea-level (km)	Tropopause height (km)	Mass of SO_2 (Mt), TOMS	Mass of SO_2 (Mt), petrological estimate	Increase in stratospheric optical depth
Sierra Negra	13/11/79	2.43[1]	14	16	1.2		0.0005
Nyamuragira	30/1/80	0.23[2]		16	0.2		
Mount St Helens	18/5/80	0.71[3]	16[18]	11.5[18]	1	0.08[22]	0.0023
Hekla	17/8/80	0.034[4]	15	8	0.5		
Ulawun	6/10/80	0.044[5]	20	16	0.2		0.0002
Alaid	27/4/81		15	12	1.1		0.0012
Pagan	15/5/81	0.5[6]	16	16	0.3		
Nyamuragira	25/12/81	0.32[2]	8	16	4		
El Chichón	4/4/82	2.3[3]	25	16	7	0.1[23]	0.0856
Galunggung	4/82 to 1/83	0.55[7]	16	16	2.5[19]	0.09[7]	
Pavlof	14/11/83	0.03[8]	10	10	0.05		
Mauna Loa	3–4/84	0.55[9]			2[20]		
Krafla	9/84	0.28[10]			0.4[20]		
Ruiz	13/11/85	0.07[11]	31	16	0.7	0.092[11]	0.0061
Nyamuragira	16/7/86	0.19[2]		16	0.8		
Banda Api	9/5/88	0.025[12]	16	16	0.2		
Redoubt	14/12/89	0.0375[13]	13	9	0.2	0.021[13]	0.0021
Pinatubo	15/6/91	10[14]	30	16	20	0.22[14]	0.1439
Hudson	12–15/8/91	6.85[15]	18	15	3.3[21]		0.009[22]
Spurr	27/6/92	0.031[16]	14.5	8	0.2		
Spurr	18/8/92	0.036[16]	13.7	8	0.4		
Spurr	17/9/92	0.039[16]	13.9	8	0.23		
Láscar	21/4/93	0.25[17]	23	16	0.4		

Tropopause height estimated from Jakosky (1986), other data from Bluth *et al.* (1997) except for: [1]Reynolds *et al.* (1995); [2]Burt *et al.* (1994); [3]Carey and Sigurdsson (1989); [4]Plinian phase only, from data in Gronvold *et al.* (1983) recalculated according to Pyle (1989);[5]McLelland *et al.* (1989); [6]Banks *et al.* (1984); [7]de Hoog *et al.* (2001); [8]McNutt (1999); [9]Lipman and Banks (1987); [10]Rossi (1997); [11]Sigurdsson *et al.* (1990); [12]Smithsonian Institution (1988); [13]Gerlach *et al.* (1994), [14]Westrich and Gerlach (1992); [15]Scasso *et al.* (1994), [16]Neal *et al.* (1995); [17]Matthews *et al.* (1997); [18]Holasek and Self (1995); [19]Bluth *et al.* (1994) includes TOMS data for explosive degassing and COSPEC data for non-explosive degassing throughout the eruption; [20]Bluth *et al.* (1993); [21]Constantine *et al.* (2000); [22]Gerlach and McGee (1994); [23]Luhr *et al.* (1984), [22]estimated value above the large Pinatubo signal.

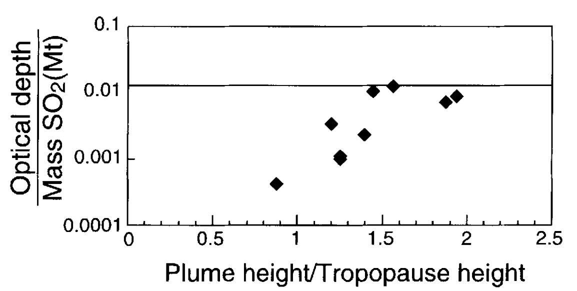

Fig. 3. Ratio of maximum increase in global stratospheric optical depth to mass of SO_2, M_{SO_2}, detected by TOMS, plotted against maximum plume height divided by tropopause height. Data and data sources from Table 1, apart for the 1991 eruption of Hudson, for which only the 15 August plume, with 2.75 Mt SO_2, is considered because its plume was higher (18 km) than that of the tropospheric 12 August plume (Constantine *et al.* 2000). The horizontal line gives the value expected from Stothers' (1984) model, assuming that all SO_2 enters the stratosphere and that there is 100% conversion to aerosol.

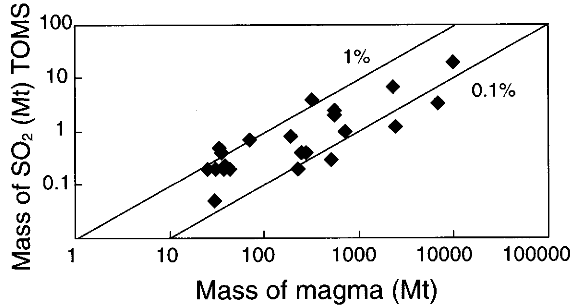

Fig. 4. TOMS data versus mass of magma erupted (data and data sources in Table 1).

The correlation in the 22 data plotted in Figure 4 can be represented by the power law

$$M_{SO_2}\text{ (in Mt)} = 1.77\ (M_{\text{magma}}\text{ (in Gt)})^{0.64} \quad (3)$$

with r^2=0.67. This power law is similar to the proportionality between M_{SO_2} and $VEI^{0.75}$ given by Pyle *et al.* (1996) for eruptions with $4 \leq VEI \leq 6$. At a first-order level, magma mass can be used as a proxy for the mass of SO_2 released.

Correlation between mass of magma and summer cooling

Eruptions that reach well into the stratosphere should cause an increase in stratospheric optical depth that is proportional to M_{SO_2} (Fig. 3 and equation 2). The resultant surface cooling should therefore scale with M_{SO_2}, which is in turn correlated with the mass of magma, M_{magma} (Fig. 4). Taking all eruptions of known mass that are believed to have entered the stratosphere in the period 1400–1994, is there a correlation between the mass of magma and amount of NH summer cooling?

The amount of summer cooling in any year is taken as that year's NHD1 anomaly minus the mean of the previous four years' anomalies. This definition allows an eruption that caused the temperature to fall to be distinguished from a benign eruption that just happened to have occurred in a generally cold period. The time series of this cooling signal derived from the data of Briffa *et al.* (1998) (http://www.ngdc.noaa.gov/paleo.html) is shown in Figure 5, and again reveals the volcanic signals of 1453, 1601, 1641, 1816, and 1912. The same conclusion emerges if the change in NHD1 anomaly from the previous year is plotted.

Figure 6 plots the summer cooling signal against magma mass, taken as a proxy for mass of SO_2, for eruptions that reached the stratosphere (Table 2). Note that the Mount Parker eruption of 1641 is excluded from the plot, because there is no measurement of the mass of magma erupted. For NH eruptions occurring at high latitude in January to June, the cooling signal for the summer in the eruption year is plotted. For all other eruptions, the signal in the year following the eruption is plotted to take account of the slow northward transport of stratospheric aerosols.

Figure 6 shows that eruptions with masses <5 to 10 Gt magma (DRE <2 to 4 km^3) are associated with signals within ±2 s.d. of the mean (i.e. $\Delta T < \pm 0.23$ °C), and are therefore regarded as having had insignificant effects on NH summer temperature. In contrast, of the eight larger eruptions, all but Santa Maria 1902 and Quizapu 1932 are associated with detectable summer cooling. The large cooling signal of –4.476 associated with the Mount Parker 1641 eruption, which had a VEI of 5 (Simkin & Siebert 2000*b*) and therefore a likely mass of order 10 Gt, is consistent with this trend. The eight plotted eruptions larger than 10 Gt (>4 km^3 dense magma or >10 km^3 tephra deposit; VEI ≥5) have a mean cooling effect of about 0.35 °C, but the data are scattered. In other words, there is a threshold eruption magnitude of 10 Mt magma, or 5 to 10 Mt SO_2 (equation (3) predicts that 10 Gt magma will release about 8 Mt SO_2) required to cause detectable NH summer cooling. Eruptions above this threshold occur on average once in a century (Simkin & Siebert 2000*a*).

Conclusions and discussion

For eruptions that penetrate well into the stratosphere, the mass of SO_2 measured by TOMS, and the increase in mean global stratospheric optical depth, are correlated by the model of Stothers

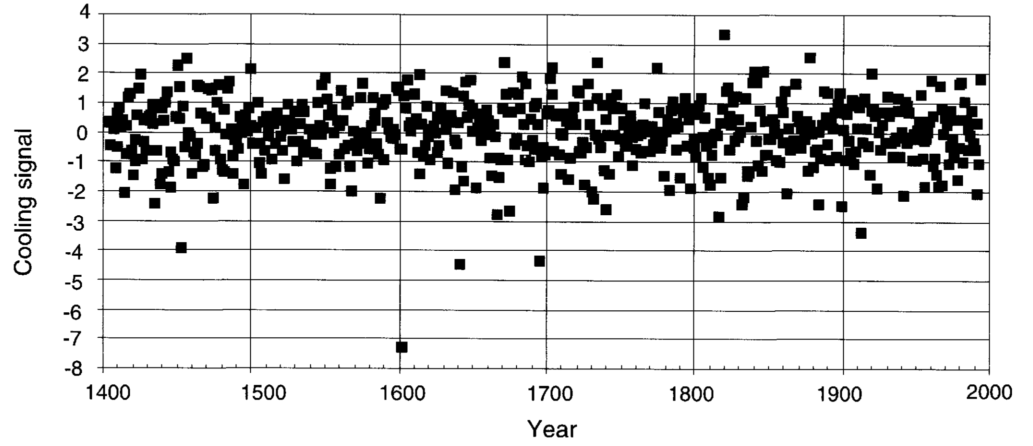

Fig. 5. The dendroclimatological record of Northern Hemisphere summer cooling in standard deviation units for each year from 1405 to 1994, calculated from the NHD1 data of Briffa *et al.* (1998) (http://www.ngdc.noaa.gov/paleo.html) using cooling signal=NHD1 value minus mean of previous four years' values. One s.d. unit represents 0.117 °C.

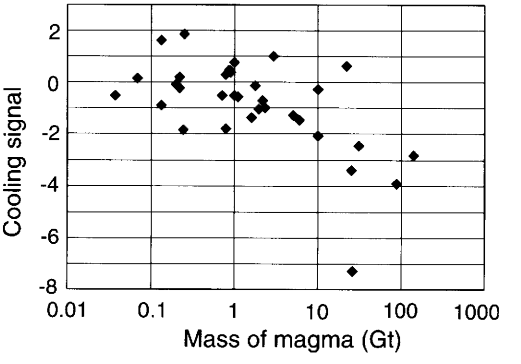

Fig. 6. Variation in cooling signal (NHD1 value minus mean of previous for years' values) with mass of erupted magma for eruptions that reached the stratosphere (data and data sources in Table 2). One s.d. unit represents 0.117 °C.

(1984). This result (Fig. 3) strengthens the argument that TOMS estimates of SO_2 released during eruptions are realistic and therefore that the petrological method frequently underestimates the S yield (Fig. 2 and Wallace 2001). The correlation between the masses of SO_2 and magma for eruptions recorded by TOMS (Fig. 2) indicates a typical yield of 0.1 to 1% SO_2 (500 to 5000 ppm S). The mass of magma erupted is therefore a guide to the mass of SO_2 released, to within half an order of magnitude.

In contrast to previous attempts to correlate SO_2 with hemispheric or global surface cooling (summarized by Sigurdsson 1990), this study does not find a simple correlation between SO_2 (or magma mass) and cooling. Instead, a threshold eruption magnitude of 10 Gt magma or equivalently 5 to 10 Mt SO_2 must be exceeded for a detectable summer cooling signal to appear. One explanation for these different conclusions is that the types of data used were different. The earlier studies used mainly the petrological method to estimate SO_2 loading and instrumental data on mean annual surface temperature. The petrological method is now recognized to sometimes be flawed, and the quality and coverage of instrumental temperature data are only sufficient for detecting significant cooling

Table 2. *Magma masses and subsequent NHD1 values (in units of standard deviation) for eruptions reaching the stratosphere.*

Eruption year	Month	Volcano	Year of NHD1 signal	Mass of magma (Gt)	NHD1 value	Change in NHD1 from from previous year	Change in NHD1 from mean of previous four years
1452		Kuwae	1453	88[1]	−4.24	−3.827	−3.914
1471	11	Sakurajima	1472	0.875[2]	−0.253	−0.471	0.468
1477	3	Veidivötn	1477	3[3]	0.403	0.814	0.998
1480		Mount St Helens	1480	5[4]	−1.299	−0.344	−1.271
1482		Mount St Helens	1482	1[4]	0.017	1.615	0.767
1510	7	Hekla	1511	0.22[5]	−0.298	−0.278	0.220
1563	6	Fogo	1563	1.1[6]	−0.585	−0.898	−0.556
1597	1	Hekla	1597	0.2[5]	−0.561	0.023	−0.092
1600	2	Huaynaputina	1601	26[7]	−6.903	−6.641	−7.308
1641		Mount Parker	1641	Unknown	−4.312	−5.084	−4.476
1663	8	Usu	1664	0.8[8]	0.483	1.262	0.298
1707	12	Fuji	1708	2.2[9]	−0.349	−1.263	−0.713
1800		Mount St Helens	1800	1[4]	−0.321	−0.335	−0.505
1815	4	Tambora	1816	140[10]	−4.326	−2.798	−2.819
1835	1	Cosiguina	1835	6[11]	−2.021	−2.732	−1.469
1875	3	Askja	1875	0.89[6]	0.167	0.529	0.381
1883	8	Krakatau	1884	31[12]	−2.887	−2.585	−2.447
1886	6	Tarawera	1887	1.8[6]	−0.109	0.931	−0.111
1902	10	Santa Maria	1903	22[6]	0.037	0.98	0.626
1907	3	Ksudach	1907	2[13]	−1.143	−1.797	−1.050
1912	6	Novarupta	1912	25[6]	−3.328	−3.873	−3.409
1932	4	Quizapu	1933	10[14]	0.565	−0.786	−0.294
1947	3	Hekla	1947	0.13[5]	0.122	−0.88	−0.905
1956	3	Bezymianny	1956	0.78[15]	−1.175	−1.548	−1.798
1963	3	Agung	1964	1.6[16]	−1.031	−0.659	−1.371
1964	11	Sheveluch	1965	0.24[17]	−2.101	−1.07	−1.824
1974	10	Fuego	1975	0.22[18]	−1.177	−0.042	−0.208
1979	4	Soufriere	1979	0.13[19]	0.01	2.814	1.641
1980	5	Mount St. Helens	1980	0.71[6]	−1.851	−1.861	−0.516
1982	4	El Chichon	1982	2.3[6]	−2.071	−2.452	−1.005
1985	11	Ruiz	1986	0.07[20]	−0.655	0.376	0.176
1989	12	Redoubt	1990	0.0375[21]	−1.027	−0.149	−0.532
1991	6	Pinatubo	1992	10[22]	−2.562	−2.338	−2.055
1993	4	Láscar	1994	0.25[23]	0.339	2.533	1.841

Data sources: NHD1 values from Briffa *et al.* (1998) (http://www.ngdc.noaa.gov/paleo.html), magma masses as follows: [1]Robin *et al.* (1994); [2]Yanagi *et al.* (1991); [3]Larsen (1984); [4]Carey *et al.* (1995); [5]Larsen *et al.* (1999); [6]Carey and Sigurdsson (1989); [7]Adams *et al.* (2001); [8]Tomiya and Takahashi (1995); [9]Miyaji *et al.* (1992); [10]Sigurdsson and Carey (1989); [11]Self *et al.* (1989); [12]Mandeville *et al.* (1996), [13]Volynets *et al.* (1999), [14]Hildreth and Drake (1992); [15]Belousov (1996); [16]Self and King (1996); [17]Belousov (1995); [18]Rose *et al.* (1978); [19]Sparks and Wilson (1982); [20]Sigurdsson *et al.* (1990); [21]Gerlach *et al.* (1994); [22]Westrich and Gerlach (1992); [23]Matthews *et al.* (1997).

signals after *c.*1860. These problems have been alleviated in the present study by using magma mass as a first-order proxy for SO_2 (as justified by the correlation shown in Fig. 4) and the long, internally consistent tree-ring series of Briffa *et al.* (1998) as a proxy for temperature. Furthermore, analyses that use the annual mean temperature combine the summer cooling and winter warming signals that can be produced by the atmospheric effects of volcanic eruptions (Robock 2000). A clearer climate signal should therefore come from treatments that consider summer cooling (or winter warming) alone. The results shown in Figure 6 indicate that a direct link between eruption magnitude and summer cooling in the Northern Hemisphere only becomes apparent for eruptions larger than 10 Gt magma that reach the stratosphere. Variation in the size of the mean cooling signal associated with eruptions above this threshold magnitude is likely to be due to uncertainties in the calibration of the tree-ring record and the variation in factors such as eruption latitude and date, aerosol production and destruction rates, and atmospheric dispersal patterns.

I sincerely acknowledge the influence of Peter Francis, whose knowledge, ideas, and dry wit are greatly missed. S. Carn, D. M Pyle, and S. Self are thanked for useful comments on this chapter.

References

ADAMS, N. K., DE SILVA, S. L., SELF, S., SALAS, G., SCHUBRING, S., PERMENTER, J. L. & ARBESMAN, K. 2001. The physical volcanology of the 1600 eruption of Huaynaputina, southern Peru. *Bulletin of Volcanology*, **62**, 493–518.

ANDRES, R. J., ROSE, W. I., KYLE, P. R., DE SILVA, S., FRANCIS, P., GARDEWEG, M. & MORENO ROA, H. 1991. Excessive sulfur dioxide emissions from Chilean volcanoes. *Journal of Volcanology and Geothermal Research*, **46**, 323–329.

ANDRONOVA, N. G., ROZANOV, E. V., YANG, F., SCHLESINGER, M. E. & STENCHIKOV, G. L. 1999. Radiative forcing by volcanic aerosols from 1850 to 1994. *Journal of Geophysical Research*, **104**, 16 807–16 826.

BANKS, N. G., KOYANAGI, R. Y., SINTON, J. M. & HONMA, K. T. 1984. The eruption of Mount Pagan Volcano, Mariana Islands, 15 May 1981. *Journal of Volcanology and Geothermal Research*, **22**, 225–269.

BELOUSOV, A. 1995. The Shiveluch volcanic eruption of 12 November 1964 – explosive eruption provoked by failure of the edifice. *Journal of Volcanology and Geothermal Research*, **66**, 357–365.

BELOUSOV, A. 1996. Deposits of the 30 March 1956 directed blast at Bezymianny volcano, Kamchatka, Russia. *Bulletin of Volcanology*, **57**, 649–662.

BLUTH, G. J. S., SCHNETZLER, C. C., KRUEGER, A. J. & WALTER, L. S. 1993. The contribution of explosive volcanism to global atmospheric sulphur dioxide concentrations. *Nature*, **366**, 327–329.

BLUTH, G. J. S., CASADEVALL, T. J., SCHNETZLER, C. C., DOIRON, S. D., WALTER, L. S., KRUEGER, A. J. & BADRUDDIN, M. 1994. Evaluation of sulfur dioxide emissions from explosive volcanism: the 1982–1983 eruptions of Galunggung, Java, Indonesia. *Journal of Volcanology and Geothermal Research*, **63**, 243–256.

BLUTH, G. J. S., ROSE, W. I., SPROD, I. E. & KRUEGER, A. J. 1997. Stratospheric loading of sulfur from explosive volcanic eruptions. *Journal of Geology*, **105**, 671–683.

BRIFFA, K. R., JONES, P. D., SCHWEINGRUBER, F. H. & OSBORN, T. J. 1998. Influence of volcanic eruptions on Northern Hemisphere summer temperature over the past 600 years. *Nature*, **393**, 450–455 (and http://www.ngdc.noaa.gov/paleo.html).

BURT, M. L., WADGE, G. & SCOTT, W. A. 1994. Simple stochastic modelling of the eruption history of a basaltic volcano: Nyamuragira, Zaire. *Bulletin of Volcanology*, **56**, 87–97.

CAREY, S. & SIGURDSSON, H. 1989. The intensity of plinian eruptions. *Bulletin of Volcanology*, **51**, 28–40.

CAREY, S., GARDNER, J. & SIGURDSSON, H. 1995. The intensity and magnitude of Holocene plinian eruptions from Mount St. Helens volcano. *Journal of Volcanology and Geothermal Research*, **66**, 185–202.

COLLING, A., DISE, N., FRANCIS, P., HARRIS, N. & WILSON, C. 1997. *The Dynamic Earth. S269 Earth and Life.* The Open University, Milton Keynes, UK.

CONSTANTINE, E. K., BLUTH, G. J. S. & ROSE, W. I. 2000. TOMS and AVHRR observations of drifting volcanic clouds from the August 1991 eruptions of Cerro Hudson. *In:* MOUGINIS-MARK, P. J., CRISP, J. A. & FINK, J. H. (eds) *Remote Sensing of Active Volcanism*. American Geophysical Union, Geophysical Monograph, **116**, 45–64.

DE HOOG, J. C. M., KOETSIER, G. W., BRONTO, S., SRIWANA, T. & VAN BERGEN, M. J. 2001. Sulfur and chlorine degassing from primitive arc magmas: temporal changes during the 1982–1983 eruptions of Galunggung (West Java, Indonesia). *Journal of Volcanology and Geothermal Research*, **108**, 55–83.

DEVINE, J. D., SIGURDSSON, H., DAVIS, A. N. & SELF, S. 1984. Estimates of sulfur and chlorine yield to the atmosphere from volcanic eruptions and potential climatic effects. *Journal of Geophysical Research*, **89**, 6309–6325.

FRANCIS, P. 1993. *Volcanoes: a Planetary Perspective*. Clarendon Press, Oxford, UK.

GERLACH, T. M. & MCGEE, K. A. 1994. Total sulfur dioxide emissions and pre-eruption vapor-saturated magma at Mount St Helens. *Geophysical Research Letters*, **21**, 2833–2836.

GERLACH, T. M., WESTRICH, H. R., CASADEVALL, T. J. & FINNEGAN, D. L. 1994. Vapor saturation and accumulation in magmas of the 1989–1990 eruption of Redoubt Volcano, Alaska. *Journal of Volcanology and Geothermal Research*, **62**, 317–337.

GRIESER, J. & SCHÖNWIESE, C.-D. 1999. Parameterization of spatio-temporal patterns of volcanic aerosol induced stratospheric optical depth and its climate radiative forcing. *Atmósfera*, **12**, 111–133.

GRONVOLD, K., LARSEN, G., EINARSSON, P., THORARINSSON, S. & SAEMUNDSSON, K. 1983. The Hekla eruption 1980–1981. *Bulletin Volcanologique*, **46**, 349–363.

HANSEN, J. & LEBEDEFF, S. 1987. Global trends of measured surface air temperature. *Journal of Geophysical Research*, **92**, 13 345–13 372.

HILDRETH, W. & DRAKE, R. E. 1992. Volcan Quizapu, Chilean Andes. *Bulletin of Volcanology*, **54**, 93–125.

HOLASEK, R. E. & SELF, S. 1995. GOES weather satellite observations and measurements of the May 18, 1980, Mount St. Helens eruption. *Journal of Geophysical Research*, **100**, 8469–8487.

JAKOSKY, B. M. 1986. Volcanoes, the stratosphere, and climate. *Journal of Volcanology and Geothermal Research*, **28**, 247–255.

JONES, P. D., BRIFFA, K. R. & SCHWEINGRUBER, F. H. 1995. Tree-ring evidence of the widespread effects of explosive volcanic eruptions. *Geophysical Research Letters*, **22**, 1333–1336.

KELLY, P. M. & SEAR, C. B. 1984. Climatic impact of explosive volcanic eruptions. *Nature*, **311**, 740–743.

KRUEGER, A. J., SCHAEFER, S. J., KROTKOV, N., BLUTH, G. & BARKER, S. 2000. Ultraviolet remote sensing of volcanic emissions. *In:* MOUGINIS-MARK, P. J., CRISP, J. A. & FINK, J. H. (eds) *Remote Sensing of Active Volcanism*. American Geophysical Union. Geophysical Monograph, **116**, 25–43.

LACIS, A., HANSEN, J. & SATO, M. 1992. Climate forcing by stratospheric aerosols. *Geophysical Research Letters*, **19**, 1607–1610.

LARSEN, G. 1984. Recent volcanic history of the Veidivötn fissure swarm, southern Iceland. *Journal of Volcanology and Geothermal Research*, **22**, 33–58.

LARSEN, G., DUGMORE, A. & NEWTON, A. 1999. Geochemistry of historical-age silicic tephras in Iceland. *The Holocene*, **9**, 463–471.

LIPMAN, P. W. & BANKS, N. G. 1987. Aa flow dynamics, Mauna Loa 1984. *United States Geological Survey Professional Paper*, **1350**, 1527–1567.

LUHR, J. F., CARMICHAEL, I. S. E. & VAREKAMP, J. C. 1984. The 1982 eruptions of El Chichon, Chiapas, Mexico: mineralogy and petrology of the anhydrite-bearing pumices. *Journal of Volcanology and Geothermal Research*, **23**, 69–108.

MCCORMICK, M. P., THOMASON, L. W. & TREPTE, C. R. 1995, Atmospheric effects of the Mt Pinatubo eruption. *Nature*, **373**, 399–404.

MCLELLAND, L., SIMKIN, T., SUMMERS, M., NIELSEN, E. & STEIN, T. C. 1989. *Global Volcanism 1975–1985*. American Geophysical Union, Washington D.C.

MCNUTT, S. 1999. Eruptions of Pavlof volcano, Alaska, and their possible modulation by ocean load and tectonic stresses: re-evaluation of the hypothesis based on new data from 1984–1998. *Pure and Applied Geophysics*, **155**, 701–712.

MANDEVILLE, C. W., CAREY, S. & SIGURDSSON, H. 1996. Magma mixing, fractional crystallization and volatile degassing during the 1883 eruption of Krakatau volcano, Indonesia. *Journal of Volcanology and Geothermal Research*, **74**, 243–274.

MASS, C. F. & PORTMANN, D. A. 1989. Major volcanic eruptions and climate: a critical evaluation. *Journal of Climate*, **2**, 566–593.

MATTHEWS, S. J., GARDEWEG, M. C. & SPARKS, R. S. J. 1997. The 1984 to 1996 cyclic activity of Lascar Volcano, northern Chile: cycles of dome growth, dome subsidence, degassing and explosive eruptions. *Bulletin of Volcanology*, **59**, 72–82.

MIYAJI, N, ENDO, K., TOGASHI, S. & UESUGI, Y. 1992. *Tephrochronological history of Mt. Fuji.* 29th IGC Field trip Guidebook Volume 4: Volcanoes and Geothermal Fields of Japan, Geological Survey of Japan, 75–109.

NEAL, C. A., MCGIMSEY, R. G., GARDNER, C. A., HARBIN, M. L. & NYE, C.J. 1995. Tephra-fall deposits from the 1992 eruptions of Crater Peak, Mount Spurr volcano, Alaska: a preliminary report on distribution, stratigraphy, and composition. *United States Geological Survey Bulletin*, **2139**, 65–79.

NEWHALL, C. G. & SELF, S. 1982. The volcanic explosivity index (VEI): an estimate of explosive magnitude for historical volcanism. *Journal of Geophysical Research*, **87**, 1231–1238.

OPPENHEIMER, C. 1996. On the role of hydrothermal systems in the transfer of volcanic sulfur to the atmosphere. *Geophysical Research Letters*, **23**, 2057–2060.

PALAIS, J. M. & SIGURDSSON, H. 1989. Petrologic evidence of volatile emissions from major historic and pre-historic volcanic eruptions. *In:* BERGER, A., DICKINSON, R. E. & KIDSON, J. W. (eds) *Understanding Climate Change*, American Geophysical Union Geophysical Monograph, **52**, IUGG Volume 7, 31–53.

PINTO, J. P., TURCO, R. P. & TOON, O. B. 1989. Self-limiting physical and chemical effects in volcanic eruption clouds. *Journal of Geophysical Research*, **94**, 11 165–11 174.

POLLACK, J. B., TOON, O. B., SAGAN, C., SUMMERS, A., BALDWIN, B. & VAN CAMP, W. 1976. Volcanic explosions and climate change: a theoretical assessment. *Journal of Geophysical Research*, **81**, 1071–1083.

PYLE, D. M. 1989. The thickness, volume and grainsize of tephra fall deposits. *Bulletin of Volcanology*, **51**, 1–15.

PYLE, D. M. 1998. How did the summer go? *Nature*, **393**, 415–416.

PYLE, D. M., BEATTIE, P. D. & BLUTH, G. J. S. 1996. Sulphur emissions to the stratosphere from explosive volcanic eruptions. *Bulletin of Volcanology*, **57**, 663–671.

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

REYNOLDS, R. W., GEIST, D. & KURZ, M.D. 1995. Physical volcanology and structural development of Sierra Negra volcano, Isabela Island, Galápagos archipelago. *Geological Society of America Bulletin*, **107**, 1398–1410.

ROBIN, C., MONZIER, M. & EISSEN, J.-P. 1994. Formation of the mid-fifteenth century Kuwae caldera (Vanuatu) by an initial hydroclastic and subsequent ignimbritic eruption. *Bulletin of Volcanology*, **56**, 170–183.

ROBOCK, A. 2000. Volcanic eruptions and climate. *Reviews of Geophysics*, **38**, 191–219.

ROBOCK, A. & MAO, J. 1995. The volcanic signal in surface temperature observations. *Journal of Climate*, **8**, 1086–1103.

ROSE JR., W. I., ANDERSON, A. T., WOODRUFF, L. G. & BONIS, S. 1978. The October 1974 basaltic tephra from Fuego Volcano: description and history of the magma body. *Journal of Volcanology and Geothermal Research*, **4**, 3–53.

ROSSI, M. J. 1997. Morphology of the 1984 open-channel lava flow at Krafla volcano, northern Iceland. *Geomorphology*, **20**, 95–112.

SADLER, J. P. & GRATTAN, J. P. 1999. Volcanoes as agents of past environmental change. *Global and Planetary Change*, **21**, 181–196.

SCAILLET, B., CLEMENTE, B., EVANS, B. & PICHAVANT, M. 1998. Redox control of sulfur degassing in silicic magmas. *Journal of Geophysical Research*, **103**, 23 937–23 949.

SCASSO, R. A., CORBELLA, H. & TIBERI, P. 1994. Sedimentological analysis of the tephra from the 12–15 August 1991 eruption of Hudson volcano. *Bulletin of Volcanology*, **56**, 121–132.

SCHNETZLER, C. C., BLUTH, G. J. S., KRUEGER, A. J. & WALTER, L. S. 1997. A proposed volcanic sulfur dioxide index (VSI). *Journal of Geophysical Research*, **102**, 20 087–20 091.

SELF, S. & KING, A. J. 1996. Petrology and sulfur and chlorine emissions of the 1963 eruption of Gunung Agung, Bali, Indonesia. *Bulletin of Volcanology*, **58**, 263–285.

SELF, S., RAMPINO, M. R. & CARR, M. J. 1989. A reappraisal of the 1835 eruption of Cosigüina and its atmospheric impact. *Bulletin of Volcanology*, **52**, 57–65.

SELF, S., ZHAO, J–X., HOLASEK, R.E., TORRES, R. C. & KING, A. J. 1996. The atmospheric impact of the 1991 Mount Pinatubo eruption. *In*: NEWHALL, C. G. & PUNONGBAYAN, R. S. (eds) *Fire and Mud: Eruptions and Lavas of Mount Pinatubo, Philippines.* University of Washington, Seattle, 1089–1115.

SIGURDSSON, H. 1990. Evidence of volcanic loading of the atmosphere and climate response. *Palaeogeography, Palaeoclimatology, Palaeoecology (Global and Planetary Change Section)*, **89**, 277–289.

SIGURDSSON, H. & CAREY, S. 1989. Plinian and co-ignimbrite tephra fall from the 1815 eruption of Tambora volcano. *Bulletin of Volcanology*, **51**, 243–270.

SIGURDSSON, H., CAREY, S., PALAIS, J. M. & DEVINE, J. 1990. Pre-eruption compositional gradients and mixing of andesite and dacite magma erupted from Nevado del Ruiz Volcano, Colombia in 1985. *Journal of Volcanology and Geothermal Research*, **41**, 127–151.

SIMKIN, T. & SIEBERT, L. 2000*a*. Earth's volcanoes and eruptions: an overview. *In:* SIGURDSSON, H., HOUGHTON, B. F., MCNUTT, S. R., RYMER, H. & STIX, J. (eds) *Encyclopedia of Volcanoes.* Academic Press, London, 249–261.

SIMKIN, T. & SIEBERT, L. 2000*b*. Appendix 2: Catalog of historically active volcanoes on Earth. *In:* SIGURDSSON, H., HOUGHTON, B. F., MCNUTT, S. R., RYMER, H. & STIX, J. (eds) *Encyclopedia of Volcanoes*. Academic Press, London, 1365–1383.

SMITHSONIAN INSTITUTION, 1988. *Banda Api, Scientific Event Alert Network Bulletin*, **13:06**.

SPARKS, R. S. J. & WILSON, L. 1982. Explosive volcanic eruptions – V. Observations of plume dynamics during the 1979 Soufrière eruption, St Vincent. *Geophysical Journal of the Royal Astronomical Society*, **69**, 551–570.

STOMMEL, H. & STOMMEL, E. 1983. *Volcano Weather: the Story of 1816, the Year Without a Summer.* Seven Seas Press, Newport, Rhode Island.

STOTHERS, R. B. 1984. The great Tambora eruption in 1815 and its aftermath. *Science*, **224**, 1191–1198.

SYMONDS, R. B., ROSE, W. I., BLUTH, G. J. S. & GERLACH, T. M. 1994.Volcanic-gas studies: methods, results, and applications. *In:* CARROLL, M. R. & HOLLOWAY, J. R. (eds) *Volatiles in Magmas*. Mineralogical Society of America, Reviews in Mineralogy, **30**, 1–66.

TOMIYA, A. & TAKAHASHI, E. 1995. Reconstruction of an evolving magma chamber beneath Usu Volcano since the 1663 eruption. *Journal of Petrology*, **36**, 617–636.

TOON, O. B. & POLLACK, J. B. 1973. Physical properties of the stratospheric aerosols. *Journal of Geophysical Research*, **78**, 7051–7056.

VOLYNETS, O. N., PONOMAREVA, V. V., BRAITSEVA, O. A., MELEKESTEV, I. V. & CHEN, CH. H. 1999. Holocene eruptive history of Ksudach volcanic massif, South Kamchatka: evolution of a large magmatic chamber. *Journal of Volcanology and Geothermal Research*, **91**, 23–42.

WALLACE, P. J. 2001. Volcanic SO_2 emissions and the abundance and distribution of exsolved gas in magma bodies. *Journal of Volcanology and Geothermal Research*, **108**, 85–106.

WESTRICH, H. R. & GERLACH, T. M. 1992. Magmatic gas source for the stratospheric SO_2 cloud from the June 15, 1991, eruption of Mount Pinatubo. *Geology*, **20**, 867–870.

YANAGI, T., ICHIMARU, Y. & HIRAHARA, S. 1991. Petrochemical evidence for coupled magma chambers beneath the Sakurajima volcano, Kyushu, Japan. *Geochemical Journal*, **25**, 17–30.