Atmospheric Volcanic Loading Derived from Bipolar Ice Cores
Accounting for the Spatial Distribution of Volcanic Deposition

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Abstract

Previous studies have used small numbers of ice core records of past volcanism to represent hemispheric or global radiative forcing from volcanic stratospheric aerosols. With the largest assembly of volcanic ice core records and state-of-the-art climate model simulations of volcanic deposition, we now have a unique opportunity to investigate the effects of spatial variations on sulfate deposition and on estimates of atmospheric loading. We have combined 44 ice core records, 25 from the Greenland and 19 from the Antarctica, and GISS modelE simulations to study the spatial distribution of volcanic sulfate aerosols in the polar ice sheets. We extracted volcanic deposition signals by applying a high pass loess filter to the time series and examining peaks that exceed twice the 31-yr running median absolute deviation. Our results suggest that the distribution of volcanic sulfate aerosol follows the general precipitation pattern in both regions, indicating the important role precipitation has played in affecting the deposition pattern of volcanic aerosols. We found a similar distribution pattern for sulfate aerosols from the 1783-1784 Laki and 1815 Tambora eruptions, as well as for the total $\beta$ activity after the 1952-54 low latitude Northern Hemisphere and 1961-62 latitude high Northern Hemisphere atmospheric nuclear weapon tests. This confirms the previous assumption that the transport and deposition of nuclear bomb test debris resemble those of volcanic aerosols. We compare three techniques for estimating stratospheric aerosol loading from ice core data: radioactive deposition from nuclear bomb tests, Pinatubo sulfate deposition in nine Antarctic ice cores, and climate model simulations of volcanic sulfate transport and deposition following the 1783 Laki, 1815 Tambora, 1912 Katmai, and 1991 Pinatubo eruptions. By applying the above calibration factors to the 44 ice core records we have estimated the stratospheric sulfate aerosol loadings for the largest volcanic eruptions during the last millennium. These loadings agree fairly well with estimates based on radiation, petrology and model simulations. We also estimate the relative magnitude of
sulfate deposition compared to the mean for Greenland and Antarctica for each ice core record, which provides a guideline to evaluate the stratospheric volcanic sulfate aerosol loading calculated from a single or a few ice core records.
1. Introduction

Understanding natural causes of climate change is vital to evaluate the relative impacts of human pollution and land surface modification on climate. Of all the natural causes, volcanic eruptions and solar variation are the two most important. Robock [2000] reviewed the impacts of volcanism on the Earth’s climate system and stressed the need for a reliable record of atmospheric loading of volcanic sulfate aerosol to be able to evaluate the causes of climatic change over the last couple of millennia. Acidity and actual sulfate records from polar firn and ice cores have provided unique details about the nature, timing, and magnitude of volcanic eruptions on regional to global scales. In the last decade, several studies [Robock and Free, 1995, 1996; Zielinski, 1995; Clausen et al., 1997; Crowley, 2000] have used one or a few ice cores to reconstruct the history of explosive volcanism. As more ice cores became available, studies have found large spatial variations in the volcanic sulfate deposition among the Greenland and Antarctic ice cores. For example, the Tambora eruption in 1815 in Indonesia has sulfate fluxes estimated to be between 22.4 kg/km^2 at Plateau Remote [Cole-Dai et al., 2000] and 133 kg/km^2 at Siple Station [Cole-Dai et al., 1997]; the 1783-1784 Laki deposition ranges from 79.7 kg/km^2 at Humboldt (78.5°N) to 323 kg/km^2 at D3 (69.8°N) [Mosley-Thompson et al., 2003], and from 100 kg/km^2 at North C (74.6°N) to 291 kg/km^2 at Milcent (70.3°N) [Clausen and Hammer, 1988]. This large spatial variability raises the question of the reliability of previous reconstructions of atmospheric volcanic sulfate loadings based on a single or only a few ice core records. It also causes problems when comparing the relative magnitudes among different eruptions as well as comparing the same volcanic signal seen in different ice sheets.

The spatial variation of volcanic sulfate deposition may be attributed to site characteristics such as surface irregularity, temperature, wind speed, and surface elevation that can modulate the local accumulation. The variation may also be caused by local or regional
circulation patterns before and during the time of deposition as well as the different deposition mechanisms. Estimations based on the average of multiple ice core records can reduce some of the uncertainties [Robock and Free, 1995, 1996; Free and Robock, 1999; Mosley-Thompson et al., 2003]. However, the number of ice core records decreases as one goes back in time and only a few ice cores are available before 1000 A.D. The uncertainty introduced by the change of ice core availability can be reduced with knowledge of the spatial distribution pattern of volcanic sulfate aerosols.

In the present study, we have incorporated the volcanic signals derived from 44 ice cores records, 25 from Greenland and 19 from Antarctica, together with NASA Goddard Institute for Space Studies (GISS) ModelE climate model simulations to estimate the spatial distribution of volcanic sulfate aerosols in Greenland and Antarctic ice sheets and the stratospheric mass loading of the largest volcanic eruptions in the past 1000 years. This paper is organized as follows: section 2 describes the ice core database and the methods to extract the volcanic signals and calculate the sulfate deposition, section 3 summarizes the spatial distribution patterns of the largest volcanic eruptions on the Greenland and Antarctica ice sheets and their comparison with those simulated by ModelE coupled to a sulfate chemistry model, and calculations of stratospheric loadings for these eruptions using new knowledge derived from bomb tests and model results are given in section 4. As this model explicitly simulates the stratospheric volcanic loading, although imperfectly, comparisons with the observations give important constraints on the validity of both the model and the observations. Discussion and conclusions are presented in section 5. This is the first study to use ice core records to investigate the spatial distribution patterns of volcanic sulfate on regional to continental scales. Our results not only provide a guideline to reconstruct a long-term volcanic forcing index with a reduced body of ice core records, but also serve as a reference to evaluate model simulations of volcanic deposition.
2. **Ice Core Database and Volcanic Deposition Calculation Methodology**

We have selected seven major low-latitude eruptions during the last millennium, Unknown (1259), Kuwae (1452 or 1453), Unknown (1809), Tambora (1815), Krakatau (1883), Agung (1963), and Pinatubo (1991), to study the spatial pattern of the volcanic sulfate deposition in the ice cores. These events were chosen because all of them are large explosive volcanic eruptions (Volcanic Explosivity Index [Newhall and Self, 1982] \( \geq 5 \)) that have signals in almost every available ice core record. In Greenland, we only have six ice cores with original sulfate data and among the six cores only three go back before 1500 A.D. We thus decided to use the 1809 Unknown, Tambora, and Krakatau eruptions as examples for low-latitude eruptions and added Laki and Katmai (1912) to represent high-latitude eruptions. *Mosley-Thompson et al.* [2003] calculated the sulfate deposition for Laki and Tambora for six PARCA (Program for Arctic Regional Climate Assessment) ice cores; *Clausen and Hammer* [1988] also computed the deposition for the same two eruptions for 12 Greenland ice cores (CH88 hereafter). We combined these two analyses with our ice core results to obtain a more comprehensive understanding of the spatial pattern in Greenland ice cores. Table 1 lists general information and references for the 26 ice core records analyzed in this study. Information about the six PARCA and 12 CH88 Greenland ice cores can be found in *Mosley-Thompson et al.* [2003] and *Clausen and Hammer* [1988] respectively. For our 26 records we first extracted volcanic deposition signals by applying a high pass loess filter to the time series and examining peaks that exceed twice the 31-yr running median absolute deviation [Gao *et al.*, 2006]. Then we evaluated the sulfate fluxes for the above eight events from the 26 cores and adjusted the timing of the signals so that the peak deposition of each event corresponds to the same year. After that, we calculated the total deposition for each eruption in the individual ice core. The resulting sulfate deposition, together with those from six PARCA and 12 CH88 records, were plotted for individual eruptions.
3. **Spatial Distribution of Volcanic Sulfate Deposition**

3.1. **Greenland Ice Cores**

Figure 1 shows the spatial distributions of the Laki, 1809 Unknown, Tambora and Krakatau volcanic sulfate deposition in the Greenland ice cores; Fig. 2 shows the spatial patterns of the deposition interpolated into 0.5°×0.5° grid points. From the figures we see that despite some local variations among the ice cores there are some common spatial patterns at regional scale, which generally follows the pattern of annual total precipitation (solid + liquid) [Fig. 6 in *Box et al.*, 2004]. For example, there is above average deposition along the intermediate elevations on the western slope with the maximum located near 70°N, below average deposition on the northeast side of the Greenland ice divide with minima found in the interior area of northern Greenland, and large deposition on the west coast of north Greenland above Melville Bay. We found a positive linear correlation between volcanic deposition and annual accumulation rate at the 95% confidence level for three (Laki, 1809 Unknown, and Tambora) of the four eruptions (Fig. 3). This is in agreement with the results of *Mosley-Thompson et al.* [2003], which indicates that the sulfate aerosols are more or less homogeneous in the atmosphere and the deposition in the ice sheet depends on the accumulation rates. No significant difference at regional scale was found between the spatial distribution of the Laki and Tambora fallouts, which implies that the deposition mechanism is probably the same for the low-latitude and high-latitude eruptions in Greenland.

From Figs. 1 and 2 we can also see that ice cores in northern Greenland (> 72°N) usually have less than average volcanic deposition, except for the ones on the west coast. Thus, volcanic stratospheric loadings based only on northern Greenland ice core records (e.g., B20, Humboldt, NorthGRIP1, and North Central) may very likely have been underestimated. The ice cores located at elevation 2000-2500 m on the west slope of the ice sheet (e.g., NASA-U, D2, and D3)
usually receive larger than average deposition, probably caused by precipitation enhancement due to orographic lifting. Maximum deposition was found in the cores located between 40°W and 45°W near 70°N, such as Crête sites B and D. Stratospheric aerosol mass loadings derived only from these cores may thus have been overestimated.

Larger spatial variations exist in deposition in the ice cores across Greenland, and the amplitude of this variability varies among different eruptions. For example, the spatial variability (defined as the ratio between the spatial standard deviation of sulfate deposition and the mean deposition) changes from 34-46% for high-latitude eruptions such as Katmai and Laki, to 41-48% for big low-latitude eruptions such as 1809 Unknown and Tambora, and further to 61-129% for smaller equatorial eruptions such as Krakatau and Agung. We also found that increasing the number of ice cores did not necessarily reduce the spatial variability. For instance, the variation of Laki and Tambora was larger than that of Katmai and 1809 Unknown although the number of cores available for Laki and Tambora was four times and twice as large as Katmai and 1809 Unknown, respectively. Spatial variability of the same or larger magnitude also exists at local scale. In the summit region of central Greenland we found a decrease of 53% from Crête site D to Crête in the volcanic sulfate aerosol deposition for the Laki eruption and 55% for Tambora, and a further decrease of 60% across the ice divide from Crête to Crête site E for Laki and 75% for Tambora. Clausen et al. [1988] also found that site E is located in an accumulation “shadow” area compared to the corresponding region west of the ice divide.

3.2. Antarctic Ice Cores

Figure 4 shows the spatial distributions of sulfate deposition for the 1809 Unknown, Tambora, Krakatau, and Agung eruptions respectively in the Antarctic ice cores. From this figure we can see a distribution pattern that is similar in regional scale for all of the four events: large deposition over the Antarctic Peninsula, West Antarctica and along the coast of East
Antarctica; and small values over the plateau of East Antarctic and around Dronning Maud Land and Victoria Land. This pattern is generally in agreement with the long-term accumulation distribution in Antarctica [Fig. 2 in Bromwich et al., 2004]. In the case of individual ice cores, Siple Station always has the highest deposition, followed by Dyer and the ITASE cores, all located in West Antarctica. Reusch et al. [1999] and Dixon et al. [2004] found that West Antarctica is the most active area of the continent in terms of atmospheric dynamics as it is surrounded by several large atmospheric low-pressure systems, the Amundsen Sea Low, the Weddell Sea Low and the Davis Sea Low, which serve as the primary transport mechanisms for moisture and aerosols to the West Antarctic ice sheet. The deposition at the four South Pole cores is relatively stable and is close to the continental averages. The air mass at the South Pole comes from West Antarctica, which brings moisture as well as aerosols. Law Dome usually has higher than average deposition. Plateau Remote, Dome C, DML-B32, Talos Dome and Hercules Névé always have lower than average deposition. Among the five sites, Plateau Remote and Dome C are located in one of the lowest long-term accumulation zones, with annual precipitation accumulation less than 50 mm/yr and elevation above 3000 km. These two cores also have the lowest correlation coefficients with the other sites, which indicates that the volcanic deposition in these sites is not representative of other regions of Antarctica and vice versa. Cole-Dai et al. [1997] found the important role of post-depositional redistribution in regulating the volcanic signals in Plateau Remote. On the other hand, Plateau Remote and DML_B32 are the only two among these 19 cores that have records during 0-1000 A.D. Proper adjustment is thus needed to account for the spatial difference when using these two records to construct volcanic forcing time series for the early periods.

Figure 5 shows the spatial distribution of sulfate deposition for the 1259 Unknown, Kuwae, and Pinatubo eruptions. Although the numbers of ice core records available for these
events are smaller than those for the previous three events, the spatial patterns of volcanic sulfate deposition are generally in agreement. The only exceptions are the larger-than-average deposition in Plateau Remote and Talos Dome for the Kuwae eruption. The reasons for these departures of volcanic sulfate deposition from the general pattern are yet to be investigated. It might have been caused by post-redistribution of the volcanic deposition or by different weather conditions, such as the circulation pattern at the time when the volcanic debris was deposited on the Antarctic ice sheet. Similar to the Greenland ice cores, we found a spatial variability of 44% and 48% for the 1809 Unknown and Tambora eruption separately across the Antarctic ice cores. The variability increased to 49%, 54%, and 65% for the moderate eruptions as Krakatau, Pinatubo, and Agung eruption respectively.

In summary, we found that despite some local variability in Greenland and a few discrepancies in Antarctica, the volcanic sulfate deposition obtained from these 25 Greenland and 19 Antarctic ice cores displays consistent spatial distribution patterns that resemble the general pattern of annual precipitation accumulation rates in Greenland and Antarctica, respectively. This indicates that the volcanic debris is more or less evenly distributed in the atmosphere before it reaches the surface. We also found that deposition in most individual ice cores is consistently smaller or larger than the Greenland or Antarctica mean values. On the other hand, spatial variability of about 45% was found for sulfate deposition across both the Greenland and Antarctic ice cores for large eruptions such as Tambora and 1809 Unknown, and this variability increases substantially for moderate eruptions such as Pinatubo and Agung. We also found site to site variations as large as a factor of four among nearby ice cores in Greenland. Therefore, it is important to obtain good spatial coverage of ice cores from different geographical areas to accurately estimate atmospheric volcanic sulfate loading. For the early periods, when there are only a few cores available from certain regions, the total volcanic deposition should be
carefully adjusted according to their ratios to the Greenland or Antarctica mean [Gao et al., 2006]. Finally, although the volcanic sulfate deposition presented here for Greenland is from three different studies that used three different methods to extract the volcanic peaks, the results should be robust because for volcanic eruptions as large as Laki, 1809 Unknown, and Tambora, the background variation should not affect the extraction of peaks and calculation of volcanic deposition no matter what method was used.

3.3. Comparison with GISS modelE simulations

Simulations were conducted of the volcanic aerosol transformation and distribution following the Pinatubo, Tambora, Katmai, and Laki eruptions using the GISS ModelE general circulation model coupled to a sulfur chemistry module [Oman et al., 2006]. This particular version of the model has 4ºx5º horizontal resolution and 23 vertical levels. The model has both dry and wet deposition schemes coupled to climate model processes [Koch et al., 2006]. A complete description of the model is given by Schmidt et al. [2006].

Due to the coarse horizontal resolution, the model cannot capture details of local, small-scale variations of sulfate deposition. But at regional and continental scales the model produced spatial patterns similar to those from ice core observations. Figure 6 shows the model-simulated volcanic sulfate deposition in both Greenland and Antarctica and their corresponding ice core observations for the Tambora eruption. The model produces low deposition in interior Northern Greenland, and high deposition in the Central and Southern Greenland and coastal regions. Over Antarctica it produces high deposition in West Antarctica and low deposition in the east Plateau region. Different from the ice core observations, the highest deposition in the model is found to be over the Transantarctic Mountains. All of the four modeled eruptions produce a similar deposition pattern in Greenland, as well as the overall Arctic region (Fig. 7). The spatial variation across both Antarctica and Greenland is also found to be mostly explained by variations
in the precipitation rate across these regions. On the other hand, the average deposition over the grid points where we have ice core measurements is as much as twice as large as that of ice core observations for Tambora (58.3 kg/km² vs. 78.4 kg/km² in Greenland, 58.8 kg/km² vs. 113.3 kg/km² in Antarctica). The difference in the magnitude of sulfate deposition between the model simulation and ice core observations could be caused by the difference in the initial injection of sulfur dioxide gas, the SO₂ to SO₄ conversion efficiency, the hemispheric partitioning of volcanic clouds, and the transport speed and pathways of the sulfate aerosols between the model realization and what actually happened during each eruption. Furthermore, the model’s coarse resolution and smooth topography may hinder its capability to accurately simulate the deposition; and its lack of a proper simulation of the Quasi-Biannual Oscillation may cause a noticeable impact on the aerosol distribution for the low latitude eruptions.

4. Estimation of Stratospheric Volcanic Sulfate Loading

By measuring the amount of sulfate that was deposited in ice cores, in theory one could do an inverse calculation to estimate the stratospheric loading, and then use this information to calculate the radiative forcing of the climate system [e.g., Stenchikov et al., 1998], one of the ultimate goals of our research. To do this inverse calculation one has to make simplifying assumptions about the area of deposition of the sulfate and the representativeness of deposition on ice for the total atmospheric loading. The simplest assumption would be that sulfate deposition was uniform worldwide, and one could just multiply the mass of sulfate per unit area measured in the ice by the surface area of the Earth (5.1×10⁸ km²). But our data from Kuwae [Gao et al., 2006; Table 4 here) show that the Southern Hemisphere (SH) polar deposition was twice of that of the Northern Hemisphere (NH), suggesting more total deposition in the SH. One could also assume uniform deposition in each hemisphere and that the ice core deposition was representative of the hemispheric average. However, our model simulations (e.g., Fig. 8) show
that most deposition occurs at midlatitudes (30º-60º) in each hemisphere, in regions of tropopause folds and strong stratosphere-troposphere transport along the jet stream and storm tracks.

An alternative is to make simplifying assumptions about the transport and deposition of the volcanic sulfate aerosols, and use some reference event to calibrate the stratospheric loading. Several previous studies [e.g., Clausen and Hammer, 1988; Langway et al., 1988; Zielinski, 1995] used factors derived from observations of radioactivity from nuclear bomb tests to estimate stratospheric sulfate loading, assuming a similar global distribution pattern between the radioactivity from bomb tests and sulfate injected into the atmosphere by violent volcanic events. Other studies [e.g., Cole-Dai and Mosley-Thompson, 1999] assumed a similar global transport pattern for all of the low-latitude eruptions and used the observed aerosol loading of the Pinatubo eruption and its deposition in six South Pole ice cores to calibrate atmospheric loadings of other low latitude eruptions. However, Zielinski [1995] found that the atmospheric loadings derived in the first group of studies are 2-5 times larger than those calculated from stratospheric observations for recent eruptions by Sato et al. [1993]. In the second group of studies the calibration was only done for the South Pole ice cores and the result may be very different for ice cores in other regions. In addition, the hemispheric partitioning of volcanic clouds can be quite different from the Pinatubo eruption, depending on the location of the eruption, height of the plume, location of the Intertropical Convergence Zone, phase of the Quasi-Biennial Oscillation, and stratospheric winds when the eruption took place.

Here we first re-examine the bomb-test-calculations using the up-to-date UNSCEAR [2000] report and update the Pinatubo-based-calibration factor using our extended assembly of ice core observations. Then we calculate another set of calibration factors using the Oman et al. [2006] coupled chemistry/climate model simulations and evaluate the sensitivity of these
calibration factors among the different methods. Finally we calculate the stratospheric loadings of the large volcanic eruptions during the past 1000 yr by applying these factors to the 44 ice core records.

4.1 Calculation of the calibration factors for tropical eruptions

Fig. 9 shows the spatial distribution of total $\beta$ activity (from $^{90}$Sr + $^{137}$Cs) from the 1952-54 low NH latitude (11°N, LNL) and 1961-62 high NH latitude (75°N, HNL) bomb tests in the Greenland ice sheet. Comparing Fig. 9 with Figs. 1 and 2, we can see that the two data sets have similar spatial coverage though they were collected from different ice cores, and the pattern of total $\beta$ activity resembles the general distribution pattern of volcanic sulfate aerosol. This confirms the previous assumption that the transport and deposition of bomb test debris resemble those of the volcanic aerosol. It also verifies, to a certain degree, the reliability of using bomb-test-derived factors to estimate stratospheric sulfate loadings. Clausen and Hammer [1988] used the total atmospheric fission injections from USA and USSR bomb tests taken from UNSCEAR [1982] and calculated the calibration factor ($L_B$ = the total $\beta$ activity injected into the atmosphere by the bomb tests / the total $\beta$ activity measured in Greenland ice cores) to calculate the global volcanic aerosol loadings for seven individual Greenland ice cores. Here we re-calculate the factor by using the stratospheric-partitioned fission yields from the most up-to-date report [UNSCEAR, 2000]. We only use the stratospheric portion of the fission yields because we are interested in estimating the volcanic sulfate loading in the stratosphere. The procedures for calculating the factors include: (1) calculate the stratospheric partitioned total fission yields (TFY) for the 1952-54 LNL bomb tests and 1961-62 HNL tests separately based on Table 1 of UNSCEAR [2000]; (2) calculate the corresponding total $\beta$ activity (TBA) by multiplying TFY with sum of the production rates of $^{90}$Sr (0.105 MCi of TBA/Mt of TFY) and $^{137}$Cs (0.159 MCi of TBA/Mt of TFY), and (3) calculate $L$ by dividing the TBA obtained in step 2 by the average TBA
measured in ice cores in Table 2 of Clausen and Hammer [1988]. The value for each step is listed in Table 2. The factor is $1.51 \times 10^9$ km$^2$ for the 1952-54 LNL bomb tests and $1.22 \times 10^9$ km$^2$ for the HNL ones. The value is much smaller than those derived by Clausen and Hammer [1988] for the LNL tests, but almost the same for the HNL tests. The reason is that only about 50% of the fission produced by the LNL tests was in the stratosphere whereas more than 90% of the fission produced by the HNL tests end up in the stratosphere at polar latitudes (Table 4 from [UNSCEAR 2000]).

In the SH the total $\beta$ activity from only three ice cores was measured by Clausen and Hammer [1988], which is too few to give a reliable estimate of a calibration factor given the large spatial variation in the Antarctic ice cores. However, since we have nine Antarctic ice cores that have Pinatubo signals, Law Dome, DML-B32, ITASE015, ITASE005, ITASE004, ITASE013, ITASE001, ITASE991, and SP2001c1, we can use these ice core records and satellite observations of Pinatubo sulfate aerosol loading to derive a calibration factor ($L_P = \text{the total sulfate aerosol injected into the atmosphere by Pinatubo / the average sulfate deposition measured in Antarctic ice cores}$) to calculate the global volcanic aerosol loadings for Antarctic ice cores. Previous studies [e.g., Krueger et al., 1995] found that Pinatubo injected 15-20 Mt of SO$_2$ gas into the middle stratosphere. Assuming 75% H$_2$SO$_4$:25% H$_2$O weight composition and a complete conversion of SO$_2$ to H$_2$SO$_4$ aerosols, this amount of SO$_2$ would produce 30-40 Mt of sulfate aerosol in the stratosphere. On the other hand, the average Pinatubo sulfate deposition derived from the above nine Antarctic ice cores is 14.8 kg/km$^2$, which gives $L_P$ ranging from $2.0 \times 10^9$ km$^2$ to $2.7 \times 10^9$ km$^2$.

The above calibration factors were calculated based on information derived from a single to a few events occurring at one latitude and altitude band under certain weather conditions, whereas the distribution of volcanic debris may differ significantly depending on the timing,
latitude and altitude of the volcanic injection, and the natural synoptic variability. For example, observations found that although the 1982 El Chichón (17ºN) and 1991 Pinatubo (15ºN) eruptions were only two degrees in latitude apart, the volcanic cloud was confined mostly to north of the Equator for the former while almost evenly distributed north and south of the Equator for the latter eruption [Robock, 2000]. For eruptions in the SH, the 1883 Krakatau eruption (6ºS) had more or less symmetric deposition of sulfate aerosols and the 1963 Agung eruption (8ºS) dispersed most of its aerosols in the SH. Gao et al [2006] found that the 1452 or 1453 Kuwae eruption (17ºS) also deposited about twice as much aerosol in the SH as in the NH. Therefore we propose that the low-latitude eruptions tend to disperse half to two thirds of the aerosols in the hemisphere where the eruptions take place depending on the particular distribution of the winds. With this assumption we can estimate a range of calibration factors based on Greenland and Antarctic ice core records, respectively, for low-latitude eruptions with different hemispheric partitioning. For Antarctic ice core records, since our calculations indicate that the average deposition of these nine ice core records with a Pinatubo signal is very close to that from the total 19 Antarctic cores for all of the four earlier tropical eruptions (i.e., the 1809 Unknown, Tambora, Krakatau, and Agung), and satellite observations showed a relatively even distribution of the Pinatubo clouds between NH and SH, it is reasonable to assume that the Pinatubo-observation derived factor ($L_P = 2.0-2.7 \times 10^9$ km$^2$) gives a fair representation of the calibration factor for tropical explosive eruptions with symmetric distribution. For eruptions that disperse 2/3 of the aerosols in the SH $L_P$ becomes 1.5-2.0 $\times 10^9$ km$^2$; and for eruptions that have 1/3 of the deposition in the SH $L_P$ is 2.7-3.6 $\times 10^9$ km$^2$. For Greenland ice core records, if we assume the bomb test debris had a symmetric distribution, $L_B$ is 1.5 $\times 10^9$ km$^2$ for eruptions with even hemispheric partitioning and 1.1 $\times 10^9$ km$^2$ or 2.0 $\times 10^9$ km$^2$ for eruptions that disperse 2/3 or 1/3 of the aerosols in the NH, respectively. On the other hand, if we assume the bomb test
debris had a 2:1 NH:SH distribution, then $L_B$ becomes $2.0 \times 10^9$ km$^2$ for eruptions with even hemispheric partitioning and either $1.5 \times 10^9$ km$^2$ or $3.0 \times 10^9$ km$^2$ for eruptions that disperse 2/3 or 1/3 of the aerosols in the NH correspondingly. Therefore, when using volcanic deposition calculated from Greenland ice core records we obtained $L_B$ ranges of $1.1-1.5 \times 10^9$ km$^2$ for eruptions that disperse 2/3 of the aerosols in the NH, of $1.5-2.0 \times 10^9$ km$^2$ for eruptions with symmetric distribution, and of $2.0-3.0 \times 10^9$ km$^2$ for eruptions that disperse 1/3 of the aerosols in the NH. Since the LNL bomb tests took place at 11ºN and were more likely to disperse more debris into the NH, the calibration factor from the high end may give more accurate estimations of the actual loadings.

The above calculations point to a mean calibration factor ($\bar{L}$) of about $2 \times 10^9$ km$^2 \pm 1 \times 10^9$ km$^2$ to be applied to deposition in each ice sheet to estimate the global aerosol loading for tropical eruptions. The uncertainty is about 50% of the mean value, which accounts mostly for the different hemispheric partitioning of the volcanic debris and in part for the uncertainty in the satellite measurement of Pinatubo atmospheric loading. If we apply half of the value ($\bar{L}_{1/2} = 1 \times 10^9$ km$^2$) to the average sulfate deposition in Greenland and Antarctica separately to calculate the loading in each hemisphere and add the two hemispheric loadings to obtain the global atmospheric loading, it is not necessary to know a priori what the hemispheric partitioning was of the initial aerosol cloud, and the ice core data will reflect the actual atmospheric loading. In this way, the uncertainty may be reduced. This assumes that removal and transport processes are on the average the same in each hemisphere, but this assumption requires further investigation with detailed validated models and observations.

The GISS modelE simulation of the 1991 Pinatubo eruption produced an average sulfate deposition of 38.4 kg/km$^2$ over Antarctica and 42.6 kg/km$^2$ over Greenland. These two average values were calculated from an area-weighted average from 70ºS to the South Pole and for
66°N-82°N, 50°W-35°W, respectively. The model simulation converted 20 Mt of SO2 gas into a sulfate aerosol yield of 36 Mt by assuming a 75 wt% H2SO4 and 25 wt% H2O composition. This gives a calibration factor ($L_{GISS}$ = the global sulfate aerosol yield in the model / the Greenland or Antarctic average sulfate deposition simulated in the model) of $0.94 \times 10^9$ km$^2$ and $0.85 \times 10^9$ km$^2$ for determining the sulfate aerosol yield from Antarctica and Greenland sulfate deposition for tropical eruptions, respectively. In the simulation of the 1815 Tambora eruption, 55 Mt of SO2 gas was injected into the 24-32 km layer which was converted into 107 Mt of sulfate aerosols. The model produced average deposition of 113.3 kg/km$^2$ and 78.4 kg/km$^2$ over the same areas in Antarctica and Greenland as for the Pinatubo eruption. Therefore, $L_{GISS}$ is $0.94 \times 10^9$ km$^2$ and $1.36 \times 10^9$ km$^2$ correspondingly. The hemispheric difference in the average deposition, and thus the calibration factors, is caused by the model’s hemispheric partitioning of the aerosol (64% in NH and 36% in SH for Pinatubo and 35% in NH and 65% in SH for Tambora). Accounting for the effect of hemispheric partitioning, we obtained the global mean calibration factor ($\overline{L}_{GISS}$) as $0.91 \times 10^9$ km$^2$ and $1.09 \times 10^9$ km$^2$ for Pinatubo and Tambora, respectively.

### 4.2 Calculation of the calibration factors for high latitude eruptions

Two high-latitude simulations, for the 1912 Katmai and the 1783-1784 Laki eruptions, were conducted using the same model. For Katmai, 5 Mt of SO2 gas was converted to 9.3 Mt of sulfate aerosol and the model produced an average sulfate deposition of 17 kg/km$^2$ over the same area of Greenland as described for Pinatubo. Therefore, $L_{GISS}$ of $0.55 \times 10^9$ km$^2$ was derived for estimating the total Katmai atmospheric loading based on sulfate deposition in Greenland. A similar calibration factor was calculated for Laki. The Laki eruption was simulated using the SO2 emission estimate from Thordarson and Self [2003], in which a total of 122 Mt of SO2 gas was injected over an 8-month period with approximately 80% going into the upper troposphere/lower stratosphere. Model simulations produce a total sulfate aerosol yield of 165
Mt over the entire eruption. The average sulfate deposition over Greenland was 284 kg/km², giving \( L_{\text{GISS}} \) as \( 0.58 \times 10^9 \) km². This value is very close to what we derived from Katmai even though the two eruptions are very different in both the duration and height of gas injection and the relative distance from Greenland. Figure 7 shows that the model produced similar spatial deposition patterns for Laki and Katmai, as well as the two tropical eruptions. Thus, we conclude that most Laki sulfate aerosols circulated around the Arctic with the polar vortex before being deposited on the Greenland ice sheet. Since the Laki and Katmai eruptions represent the breadth of different types of NH high latitude eruptions it is reasonable to assume that the average calibration factor of the two events (\( 0.57 \times 10^9 \) km²) is applicable to all NH high latitude eruptions.

4.3 Estimates of stratospheric volcanic aerosol loadings

Table 3 summarizes the calibration factors calculated from the above three methods, from which we find that the model-derived factors are substantially smaller than those from the other two methods. For tropical eruptions, it is very likely that the model underestimates the factors by up to 50% due to the faster transport of volcanic aerosols from tropical eruptions which resulted in a greater sulfate deposition in high latitudes. The bomb-test calculation and Pinatubo observation derived factors may thus provide better estimates of the stratospheric mass loading of the tropical eruptions. No matter what method was used, the resulting calibration factor is significantly larger than the total area of the Earth (\( 0.51 \times 10^9 \) km²). Our model simulations (Fig. 8) suggest that this is because most of the deposition is in the midlatitudes (30°-60°) in each hemisphere.

For the NH high latitude eruptions, since we have no candidate eruption that has both satellite observations and an ice core sulfate record, we cannot derive a calibration factor as we did for Pinatubo. The bomb test calculation is also not a reliable method because most HNL
Bomb tests were conducted in Novaya Zemlya (75°N, 60°E), 11°-17° north of where the Laki (64°N) and Katmai (58°N) eruptions took place, as well as 8° north of the Arctic Circle. Therefore, the transport and deposition of the nuclear debris may not be a good model for that from lower latitude volcanic eruptions. In addition, the HNL bomb tests were characterized by near instantaneous release of volatiles to heights larger than 20 km, whereas Laki in particular featured an eight-month long eruption with 13-14 km high plumes and the atmospheric mass loading was confined to the low stratosphere and troposphere [Thordarson et al., 1993; Fiacco et al., 1994; Thordarson et al., 2001]. Since both the atmospheric circulation and lifetime of aerosols are fundamentally different for the stratosphere and troposphere [Holton et al., 1995], the HNL bomb tests likely misrepresent the transport and deposition of NH high latitude volcanic aerosols. Model simulations of the Laki and Katmai eruption, on the other hand, were found to give reasonable estimates of the transport and dispersal of the aerosols [Oman et al., 2006].

According to the previous discussion, we decided to combine the bomb test calculation and Pinatubo observation derived factors to calculate the stratospheric sulfate aerosol loadings for tropical eruptions, while using the model-derived calibration factors to calculate the stratospheric sulfate aerosol loadings for high-latitude eruptions. Specifically for the tropical eruptions we applied $\overline{L}_{1/2} = 1.0 \times 10^9$ km$^2$ to the Greenland and Antarctic average deposition separately to calibrate the loadings for the corresponding hemisphere, then added the loading calculated from each hemisphere to give the final estimate of the global loading; for NH high latitude eruptions we applied $\overline{L}_{GISL} = 0.57 \times 10^9$ km$^2$ to the average deposition in Greenland ice cores and obtained the NH aerosol loading which also stands for the global loading. Table 4 lists the atmospheric sulfate loadings of the largest eruptions during the past 1000 years obtained in this way.
As shown in Table 4 we estimated the stratospheric loading of Tambora sulfate aerosols to be 115 Tg. This value is very close to the estimation of 93-118 Tg based on a petrology study [Self et al., 2004] but substantially smaller than the estimation of ~200 Tg by Stothers [1984]. The latter is probably overestimated due to a local dense region of stratospheric aerosols [Self et al., 2004]. The stratospheric loading of Laki is 88 Tg of sulfate aerosols, which is smaller than the estimates based on radiation [200 Mt, Stothers, 1996] and geology [200 Mt, Thordarson and Self, 2003], the second of which was used as input for the climate model simulation of Oman et al. [2006]. This is because we only estimated the stratosphere loading while the other two studies reported the total atmospheric loading instead of the stratospheric component. Observational studies [Thordarson et al., 1993; Fiacco et al., 1994] found that about one third to one half of the emissions from Laki were injected into the stratosphere. Thus if we multiply our estimates of Laki loading by a factor of two, this would give a total atmospheric loading of 176 Mt which is in line with other estimates.

4.4 Improvements from previous loading estimates and remaining uncertainties

The above stratospheric volcanic mass loadings are the best estimations we could obtain with the available information, albeit imperfect. Robock and Free [1995] identified eight problems in using ice core records as measures of volcanic aerosol loading: (1) other sources of acid and bases; (2) other sources of sulfate; (3) dating uncertainties; (4) local volcanoes; (5) limited knowledge of the aerosol’s pathway from the stratosphere to the ice; (6) stochastic nature of snowfall and dry deposition; (7) mixing due to blowing snow; (8) temperature dependence of ECM measurements. In the present study we used only the ice core sulfate records, which eliminated the first and the last problems. The signal extraction and deposition calculation methodology minimized the errors associated with problem (2) and (3); and our extended body of ice cores from both Greenland and Antarctica helped to distinguish local volcanoes from
tropical ones in conjunction with the recent VEI index [Siebert and Simkin, 2002]. The information of the spatial distribution pattern obtained in section 3 was used to estimate and reduce the uncertainties caused by problems (6) and (7). The remaining challenge is to find out the relationship between atmospheric aerosol loadings and the amount of sulfate deposited in the snow and ice we were measuring. While the distribution of volcanic debris differs significantly depending on the latitude, altitude, timing of the eruption and the direction of the winds when the eruption took place, the calibration factors used in this study were derived from a single to a few events occurring at one latitude and altitude band under certain weather conditions. To account for this limitation, we calculated $L_B$ and $L_P$ under different assumptions of hemispheric partitioning and estimated the uncertainty range for the mean calibration factor $\bar{L}$. Simultaneous use of ice core records from both poles tells something about the hemispheric partitioning of volcanic clouds and thus would be an optimal way to estimate the global loadings. However, fewer ice core observations are available as one goes back in time to give a reliable indication of the hemispheric partitioning as well as the magnitude of the atmospheric mass loading. Therefore, proper adjustments are needed when using few ice cores to estimate strength of the earlier eruptions or the results should be interpreted with caution.

Furthermore, the bomb-test-calculation and Pinatubo-observation derived factors are only for NH and SH respectively, due to the limited radioactivity and Pinatubo deposition measurements. In addition, calibration against a well-known eruption requires a linear relationship between the sulfate that emitted into the stratosphere and that recovered from the ice core measurement. The linear relationship does not always hold, especially for the large eruptions.

Our state-of-the-art climate model produces twice the deposition as ice core observations for both Tambora and Pinatubo, and hence the calibration factor is about half that derived from
observations, assuming the model used the correct atmospheric injections of SO₂, but the model was not developed for this purpose. The general pattern of sulfate deposition shown in Fig. 8, seems reasonable, but a model with much more accurate and detailed stratospheric circulation and stratospheric-tropospheric exchange needs to be developed and applied to this problem. Such a model should include high-resolution tropospheric components over the ice sheets to accurately simulate the dry and wet deposition processed there, including the detailed effects of orography.

5. Conclusions

We have used 44 ice core records, 25 from Greenland and 19 from Antarctica, and GISS modelE simulations to examine the spatial distribution of the volcanic sulfate aerosol in the polar ice sheets. The model results and ice core observations are in general agreement, both of which suggest that the distribution of volcanic sulfate aerosol follows the general precipitation pattern in both regions, indicating the important role precipitation has played in removing the stratospheric volcanic aerosols. We estimated the relative magnitude of sulfate deposition associated with individual ice cores with respect to the Greenland or Antarctic means and these results provides a guideline to at least qualitatively evaluate stratospheric volcanic sulfate aerosol loading based on single or a few ice core records. We found a similar spatial distribution pattern between the volcanic sulfate deposition and the bomb test fallout in the Greenland ice cores the low-latitude and high-latitude volcanic debris, which confirms the previous assumption that the transport and deposition of bomb test debris resemble those of the volcanic aerosol. We compare three techniques for estimating stratospheric aerosol loading from ice core data, radioactive deposition from nuclear bomb tests, Pinatubo sulfate deposition in nine Antarctic ice cores, and climate model simulations of volcanic sulfate transport and deposition following the 1783 Laki, 1815 Tambora, 1912 Katmai, and 1991 Pinatubo eruptions. Combining these calibration factors
with the adjustments accounting for the spatial variation, we produce estimates of the stratospheric sulfate aerosol loadings for large volcanic events during the past millennia. The results lie in the middle of the estimations based on petrology, radiation geology, and model studies and can be used to calculate the radiative forcing for the aerosol clouds and effects of these volcanic eruptions on climate.

**Acknowledgments.** We thank all the scientists who have supplied us with ice core records both for the difficult work of obtaining and analyzing the cores and for allowing us to use them. This work is supported by NOAA grant NA03-OAR-4310155 and NSF grant ATM-0313592. Model development and computer time at GISS supported by NASA climate modeling grants.
References


Delmas, R. J., S. Kirchner, J. M. Palais, and J. R. Petit (1992), 1000 years of explosive volcanism recorded at the South-Pole, *Tellus B*, 44, 335-350.


Table 1. Ice core time series used in the study. DEP = dielectric profiling, NSS \( \text{SO}_4 \) = non-sea-salt sulfate, CFA = continuous flow analysis, NSS-conductivity = non-sea-salt conductivity, EXS = excess sulfate.<p></p>

<table>
<thead>
<tr>
<th>Name</th>
<th>Location</th>
<th>Period</th>
<th>Resolution</th>
<th>Measure Type</th>
<th>Type</th>
<th>Units</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>NGT_B20</td>
<td>79ºN, 36.5ºW</td>
<td>830-1993</td>
<td>12/yr</td>
<td>CFA</td>
<td>ng/g (ppb)</td>
<td></td>
<td>Bigler et al. [2002]</td>
</tr>
<tr>
<td>NorthGRIP1</td>
<td>75.1ºN, 42.3ºW</td>
<td>190-1969</td>
<td>1/yr</td>
<td>total ( \text{SO}_4 )</td>
<td>µequiv./kg</td>
<td></td>
<td>Gao et al. [2006]</td>
</tr>
<tr>
<td>GISP2*</td>
<td>72.6ºN, 38.5ºW</td>
<td>1-1984</td>
<td>0.5/yr</td>
<td>NSS ( \text{SO}_4 )</td>
<td>ppb</td>
<td></td>
<td>Zielinski [1995]</td>
</tr>
<tr>
<td>Greenland Site T</td>
<td>72.6ºN, 38.5ºW</td>
<td>1731-1989</td>
<td>1/yr</td>
<td>EXS</td>
<td>kg/km²</td>
<td></td>
<td>Mosley-Thompson et al. [1993]</td>
</tr>
<tr>
<td>Greenland Site A</td>
<td>70.8ºN, 36ºW</td>
<td>1715-1985</td>
<td>1/yr</td>
<td>EXS</td>
<td>kg/km²</td>
<td></td>
<td>Mosley-Thompson et al. [1993]</td>
</tr>
<tr>
<td>20D*</td>
<td>65ºN, 45ºW</td>
<td>1767-1983</td>
<td>1/yr</td>
<td>NSS ( \text{SO}_4 )</td>
<td>ng/g</td>
<td></td>
<td>Mayewski et al. [1990]</td>
</tr>
<tr>
<td>Mt. Logan*</td>
<td>60.6ºN, 140.6ºW</td>
<td>1689-1979</td>
<td>1/yr</td>
<td>total ( \text{SO}_4 )</td>
<td>µequiv./l</td>
<td></td>
<td>Mayewski et al. [1993]</td>
</tr>
<tr>
<td>Law Dome</td>
<td>66.7ºS, 112.8ºE</td>
<td>1301-1995</td>
<td>12/yr</td>
<td>NSS ( \text{SO}_4 )</td>
<td>µequiv./l</td>
<td></td>
<td>Palmer et al. [2002]</td>
</tr>
<tr>
<td>Dyer</td>
<td>70.7ºS, 65ºW</td>
<td>1505-1989</td>
<td>1/yr</td>
<td>total ( \text{SO}_4 ) flux</td>
<td>kg/km²</td>
<td></td>
<td>Cole-Dai et al. [1997]</td>
</tr>
<tr>
<td>G15*</td>
<td>71.2ºS, 46ºE</td>
<td>1210-1983</td>
<td>varies</td>
<td>DEP</td>
<td>µS/m</td>
<td></td>
<td>Moore et al. [1991]</td>
</tr>
<tr>
<td>Talos Dome</td>
<td>72.8ºS, 159.1ºE</td>
<td>1217-1996</td>
<td>varies</td>
<td>NSS ( \text{SO}_4 )</td>
<td>µequiv./l</td>
<td></td>
<td>Stenni et al. [2002]</td>
</tr>
<tr>
<td>Hercules Névêrie</td>
<td>73.1ºS, 165.5ºE</td>
<td>1774-1992</td>
<td>1/yr</td>
<td>NSS ( \text{SO}_4 )</td>
<td>µequiv./l</td>
<td></td>
<td>Stenni et al. [2002]</td>
</tr>
<tr>
<td>Dome C*</td>
<td>74.7ºS, 124.2ºE</td>
<td>1763-1973</td>
<td>1/yr</td>
<td>NSS ( \text{SO}_4 )</td>
<td>µequiv./l</td>
<td></td>
<td>Legrand and Delmas [1987]</td>
</tr>
<tr>
<td>DML_B32</td>
<td>75ºS, 0ºW</td>
<td>159-1997</td>
<td>varies</td>
<td>NSS ( \text{SO}_4 )</td>
<td>ng/g</td>
<td></td>
<td>Traufetter et al. [2004]</td>
</tr>
<tr>
<td>Siple Station</td>
<td>76ºS, 84.3ºW</td>
<td>1417-1983</td>
<td>1/yr</td>
<td>Total ( \text{SO}_4 ) flux</td>
<td>kg/km²</td>
<td></td>
<td>Cole-Dai et al. [1997]</td>
</tr>
<tr>
<td>ITASE 01-5</td>
<td>77ºS, 89ºW</td>
<td>1781-2002</td>
<td>varies</td>
<td>( \text{SO}_4 )</td>
<td>µg/l</td>
<td></td>
<td>Dixon et al. [2004]</td>
</tr>
<tr>
<td>ITASE 00-5</td>
<td>77.7ºS, 124ºW</td>
<td>1708-2001</td>
<td>varies</td>
<td>( \text{SO}_4 )</td>
<td>µg/l</td>
<td></td>
<td>Dixon et al. [2004]</td>
</tr>
<tr>
<td>ITASE 00-4</td>
<td>78ºS, 120ºW</td>
<td>1799-2001</td>
<td>varies</td>
<td>( \text{SO}_4 )</td>
<td>µg/l</td>
<td></td>
<td>Dixon et al. [2004]</td>
</tr>
<tr>
<td>ITASE 01-3</td>
<td>78.1ºS, 95.6ºW</td>
<td>1859-2002</td>
<td>varies</td>
<td>( \text{SO}_4 )</td>
<td>µg/l</td>
<td></td>
<td>Dixon et al. [2004]</td>
</tr>
<tr>
<td>ITASE 00-1</td>
<td>79.4ºS, 111ºW</td>
<td>1651-2001</td>
<td>varies</td>
<td>( \text{SO}_4 )</td>
<td>µg/l</td>
<td></td>
<td>Dixon et al. [2004]</td>
</tr>
<tr>
<td>ITASE 99-1</td>
<td>80.6ºS, 122.6ºW</td>
<td>1713-2000</td>
<td>varies</td>
<td>( \text{SO}_4 )</td>
<td>µg/l</td>
<td></td>
<td>Dixon et al. [2004]</td>
</tr>
<tr>
<td>Plateau Remote*</td>
<td>84ºS, 43ºE</td>
<td>1-1986</td>
<td>1/yr</td>
<td>( \text{SO}_4 )</td>
<td>ppb</td>
<td></td>
<td>Cole-Dai et al. [2000]</td>
</tr>
<tr>
<td>PS1*</td>
<td>90ºS</td>
<td>1010-1984</td>
<td>1/yr</td>
<td>NSS ( \text{SO}_4 )</td>
<td>ng/g</td>
<td></td>
<td>Delmas et al. [1992]</td>
</tr>
<tr>
<td>PS14*</td>
<td>90ºS</td>
<td>1800-1984</td>
<td>1/yr</td>
<td>NSS ( \text{SO}_4 )</td>
<td>ng/g</td>
<td></td>
<td>Delmas et al. [1992]</td>
</tr>
<tr>
<td>SP2001 (core 1)</td>
<td>90ºS</td>
<td>905-1999</td>
<td>1/yr</td>
<td>Total ( \text{SO}_4 ) flux</td>
<td>kg/km²</td>
<td></td>
<td>Budner and Cole-Dai [2003]</td>
</tr>
<tr>
<td>SP95</td>
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<td>1487-1992</td>
<td>varies</td>
<td>( \text{SO}_4 )</td>
<td>µg/l</td>
<td></td>
<td>Dixon et al. [2004]</td>
</tr>
</tbody>
</table>

*Cores used by Robock and Free [1995].
**Table 2.** Calculation of stratospheric partitioning of total β activity and the calibration factors. LNL is Low Northern Hemisphere Latitude and HNL is High Northern Hemisphere Latitude.

<table>
<thead>
<tr>
<th>Years of β deposition</th>
<th>Stratospheric partitioning of total fission yield (Mt)a</th>
<th>Total β activity injected in the stratosphere (MCi)b</th>
<th>Average total β activity in Greenland ice cores (mCi/km²)c</th>
<th>Calibration factor (L_B) for Greenland ice core records (×10⁹ km²)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1952-54 (LNL)</td>
<td>18.11</td>
<td>4.78</td>
<td>3.17</td>
<td>1.51</td>
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<tr>
<td>1961-62 (HNL)</td>
<td>69.64</td>
<td>18.39</td>
<td>15.08</td>
<td>1.22</td>
</tr>
</tbody>
</table>

aFrom Table 1 of UNSCEAR [2000].
bBy applying the production rate of ⁹⁰Sr and ¹³⁷Cs of 0.105 MCi/Mt and 0.159 MCi/Mt.
cFrom Table 2 of Clausen and Hammer [1988].

**Table 3.** Calibration factors derived from three different methods (×10⁹ km²).

<table>
<thead>
<tr>
<th>Method</th>
<th>For tropical eruptions based on NH ice cores</th>
<th>For tropical eruptions based on SH ice cores</th>
<th>For NH high latitude eruptions based on NH ice cores</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>2:1 NH vs. SH</td>
<td>1:1 NH vs. SH</td>
<td>1:2 SH vs. NH</td>
</tr>
<tr>
<td>Bomb test calculation (L_B)</td>
<td>1.1 – 1.5</td>
<td>1.5 – 2.0</td>
<td>2.0 – 3.0</td>
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<td>Pinatubo observations (L_P)</td>
<td>N/A</td>
<td>N/A</td>
<td>N/A</td>
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<tr>
<td>Climate model simulations (L_GISS)</td>
<td>N/A</td>
<td>N/A</td>
<td>N/A</td>
</tr>
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</table>
Table 4. Estimates of stratospheric sulfate loading from large explosive volcanic eruptions.

<table>
<thead>
<tr>
<th>Eruption</th>
<th>Year</th>
<th>Number of cores used in average</th>
<th>Ice core average sulfate deposition (kg/km²)</th>
<th>Calibration factor ($\times 10^9$ km²)</th>
<th>Hemispheric stratospheric sulfate aerosol loading (Tg)</th>
<th>Global stratospheric sulfate aerosol loading (Tg)</th>
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<tbody>
<tr>
<td>Unknown</td>
<td>1259</td>
<td>NH (3)</td>
<td>145.9*</td>
<td>1</td>
<td>145.9</td>
<td>258</td>
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<tr>
<td></td>
<td></td>
<td>SH (5)</td>
<td>112*</td>
<td>1</td>
<td>112</td>
<td></td>
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<tr>
<td>Kuwae</td>
<td>1452</td>
<td>NH (2)</td>
<td>44.6**</td>
<td>1</td>
<td>44.6</td>
<td>138</td>
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<td></td>
<td>SH (7)</td>
<td>92.9**</td>
<td>1</td>
<td>92.9</td>
<td></td>
</tr>
<tr>
<td>Laki</td>
<td>1783</td>
<td>NH (24)</td>
<td>154.3</td>
<td>0.57</td>
<td>88</td>
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<tr>
<td>Unknown</td>
<td>1809</td>
<td>NH (11)</td>
<td>29.2</td>
<td>1</td>
<td>29.2</td>
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<tr>
<td></td>
<td></td>
<td>SH (17)</td>
<td>27.1</td>
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<td>27.1</td>
<td></td>
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<tr>
<td>Tambora</td>
<td>1815</td>
<td>NH (22)</td>
<td>58.3</td>
<td>1</td>
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<td>SH (17)</td>
<td>57.0</td>
<td>1</td>
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<tr>
<td>Krakatau</td>
<td>1883</td>
<td>NH (7)</td>
<td>13.1</td>
<td>1</td>
<td>13.1</td>
<td>25</td>
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<td>SH (18)</td>
<td>12.1</td>
<td>1</td>
<td>12.1</td>
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<td>Katmai</td>
<td>1912</td>
<td>NH (6)</td>
<td>21.8</td>
<td>0.57</td>
<td>12.4</td>
<td>N/A</td>
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<tr>
<td>Agung</td>
<td>1963</td>
<td>NH (7)</td>
<td>6.7</td>
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<td>6.7</td>
<td>17</td>
</tr>
<tr>
<td></td>
<td></td>
<td>SH (17)</td>
<td>10.3</td>
<td>1</td>
<td>10.3</td>
<td></td>
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<tr>
<td>Pinatubo</td>
<td>1991</td>
<td>SH (10)</td>
<td>14.8</td>
<td>1</td>
<td>14.8</td>
<td>30</td>
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</tbody>
</table>

* The original deposition in Greenland and Antarctic ice sheet was multiplied by 1.13 and 1.27, respectively to account for the spatial variation. The multipliers were calculated by comparing the average deposition of the total 7 Northern Hemisphere and 17 Southern Hemisphere ice cores to those of the 3 Northern Hemisphere and 7 Southern Hemisphere cores that have 1259 Unknown signal.

** The original deposition in Greenland and Antarctic ice sheet was multiplied by 1.81 and 1.02, respectively. See Gao et al. [2006] for details.
Figure 1. Spatial distribution of Laki, 1809 Unknown, Tambora, and Krakatau sulfate deposition (kg/km$^2$) in Greenland ice cores. The colors are defined so that the ones in blue indicate smaller than average deposition and the ones in yellow, orange, and red indicate larger than average.
Figure 2: Same as Fig. 1 with the deposition was interpolated into 0.5 by 0.5 degree grid points.
Figure 3. Relationship between the annual snow accumulation rates and the total sulfate fluxes in Greenland ice cores.
Figure 4. Spatial distribution of 1809 Unknown, Tambora, Krakatau and Agung sulfate deposition (kg/km²) in Antarctic ice cores. The colors are defined so that the ones in blue indicate smaller than average deposition and the ones in yellow, orange, and red indicate larger than average.
Figure 5. Same as Fig. 4 but for the 1259 Unknown, Kuwae, and Pinatubo eruption.
Figure 6. Comparison between the total Tambora sulfate deposition (kg/km²) in Greenland (top panel) and Antarctic (bottom panel) ice core observations and in the GISS simulations.
Figure 7. GISS simulated Laki, Tambora, Katmai, and Pinatubo deposition (kg/km²) in the Arctic region. The colors are defined so that the ones in blue indicate smaller than average deposition for 66°N-82°N, 50°W-35°W and the ones in yellow, orange, and red indicate larger than average.
Figure 8. Global spatial distribution of the volcanic sulfate deposition (kg/km²) after the 1815 Tambora eruption, simulated by the GISS modelE. 55 Mt of SO₂ gas was put into the 24-32 km layer, which converted into 107 Mt of sulfate aerosols assuming a 75%:25% H₂SO₄:H₂O weight composition.
Figure 9. Spatial distribution of the total $\beta$ activity from the 1952-54 LNL and 1961-62 HNL bomb tests.