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Climatic Impact of the A.D. 1783 Asama (Japan) Eruption was Minimal: Evidence from the GISP2 Ice Core

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Climatic impact of the A.D. 1783 Asama (Japan) eruption was minimal: Evidence from the GISP2 ice core

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Abstract. Assessing the climatic impact of the A.D. 1783 eruption of Mt. Asama, Japan, is complicated by the concurrent eruption of Laki, Iceland. Estimates of the stratospheric loading of H_2SO_4 for the A.D. 1108 eruption of Asama derived from the SO_4^{2-} time series in the GISP2 Greenland ice core indicate a loading of about 10.4 Tg H_2SO_4 with a resulting stratospheric optical depth of 0.087. Assuming sulfur emissions from the 1783 eruption were only one-third of the 1108 event yields a H_2SO_4 loading value of 3.5 Tg and a stratospheric optical depth of only 0.029. These results suggest minimal climatic effects in the Northern Hemisphere from the 1783 Asama eruption, thus any volcanically-induced cooling in the mid-1780s is probably due to the Laki eruption.

Introduction

Asama volcano (36.5°N, 138.9°E), Honshu, Japan, has erupted many times over the last 1000 years [*Aramaki*, 1956; 1957; 1963], but only two of these events have been large enough to definitely warrant a volcanic explosivity index (VEI) of at least 4 [*Simkin et al.*, 1981]. The earliest and most explosive event was in September 1108 (VEI = 5?) The more recent event was in May-August 1783 (VEI = 4). Despite the greater volume of erupted material from the 1108 eruption [*Aramaki*, 1963], the 1783 event is probably the best-known of these eruptions. This notoriety is especially due to its uncertain climatic effects on the Northern Hemisphere. In fact, direct evidence for the atmospheric and climatic effects of the 1783 Asama eruption is not easily obtained, because of its timing in relationship to the lengthy Icelandic eruption of Laki [June, 1783-February, 1784; *Thordarson and Self*, 1994].

It is well established both from instrumental records [*Angell and Korshover*, 1985] and from historical and proxy records [*Mikami and Tsukamura*, 1992; *Wood*, 1992] that climatic conditions during the mid-1780s were much cooler than average conditions in the late 1700s. However, questions may arise regarding the magnitude of climatic cooling due to the individual events. In particular, important information that may be used to evaluate the atmospheric effects of large fissure eruptions like Laki versus more explosive plinian eruptions like Asama is lost because of the possible cumulative effects of the two eruptions. Thus, many studies of the atmospheric effects of volcanism in the mid-1780s often lump the Laki and Asama eruptions together.

We present here a different approach to estimate the amount of stratospheric loading of the 1783 Asama eruption and ultimately its potential climatic effects. By using the amount of volcanically derived SO_4^{2-} recorded in the Greenland Ice Sheet Project Two (GISP2) ice core for the 1108 Asama eruption and the relative magnitude of the 1108 and 1783 events, we can estimate the stratospheric loading of H_2SO_4 for the 1783 Asama eruption. All ice cores recovered from Greenland are characterized by the presence of a large volcanic signal around 1783 [e.g., *Clausen and Hammer*, 1988], but an estimate of the magnitude of the signal that can be attributed to the Asama eruption versus the proximal Laki eruption has not been attempted. We will do that here as well as provide information on the atmospheric effects of the 1108 eruption.

Previous workers have approximated the amount of atmospheric loading of historical Asama eruptions through the composition of glass inclusions trapped in phenocrysts (i.e., petrologic method; [*Devine et al.*, 1984]). Sulfur emission estimates for the 1108 event were 0.33 Tg resulting in an atmospheric loading of 1 Tg H_2SO_4 by the petrologic method, whereas sulfur emission estimates for the 1783 event were 0.1 Tg resulting in an atmospheric loading of 0.3 Tg H_2SO_4 [*Kohno et al.*, 1993]. Thus, the atmospheric loading of the 1108 eruption is thought to be about three times that of the 1783 eruption [*Kohno et al.*, 1993]. Unfortunately, the petrologic technique may underestimate the amount of atmospheric loading of sulfur gases from an eruption possibly due to an unknown sulfur source (as was found after El Chichón, 1982; [*Devine et al.*, 1984]). It is thus necessary to compare the amount of atmospheric loading estimated by the petrologic method with that from other sources, like ice cores.

We analyzed the complete suite of major ions in bi-yearly samples for the GISP2 core using ion chromatography [*Mayewski et al.*, 1993]. High peaks in SO_4^{2-} and occasionally in Cl^- are thought to be representative of the deposition of volcanically-derived aerosols [e.g., *Mayewski et al.*, 1993]. Specific laboratory techniques and the protocol used to eliminate contamination given in *Mayewski et al.* [1993]. We also filtered meltwater samples in the 1783/84 section in an attempt to identify volcanic glass with a scanning electron microscope and microprobe (see *Fiacco et al.*,1993].

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Depth-age scale for the core was determined by the counting of annual layers found in the physical stratigraphy of the ice and of peaks in electrical conductivity, laser particulates, and $\delta^{18}O$. Dating error is now thought to be 1% for the last 30,000 years of record [*Meese et al.*, 1994], but this is a very conservative value and errors are probably much less than that for the last 1000 years for which we are concerned. There is an additional error of ± two years because of the biyearly sampling scheme used and, we must also consider the lag between the time of the eruption and deposition on the ice. Nevertheless, we feel that we may accurately identify the volcano responsible for most signals over the last 1000 years [*Zielinski et al.*, 1994]. Subannual sampling in some volcanic sections, like for Laki [*Fiacco et al.*, in press], have verified that the timing of the volcanic signal in the biyearly sampling is appropriate for the given eruption. Age and VEI of all eruptions discussed are from *Simkin et al.* [1981] and more recent supplements [T. Simkin and L. Siebert, personal communication].

To estimate the amount of stratospheric loading for each Asama eruption, we used the technique of *Clausen and Hammer* [1988]. *Clausen and Hammer* [1988] analyzed ice core samples across Greenland for the concentration of fallout from bomb testing in the early 1950s at low latitudes (11°N) and in the early 1960s at high latitudes (75°N). Because the source of the bomb material deposited on Greenland originated from the stratospheric injection of a known magnitude from a single point source, they assumed that the stratospheric transport of volcanic aerosols and eventual deposition on the Greenland Ice Sheet is similar to that of material from nuclear bomb testing. However the amount of bomb fallout recorded in Greenland from low latitude testing was lower than that from high latitude testing indicating the loss of aerosols during longer transport. To account for this loss a multiplier was determined to convert the concentration of deposited material on the ice sheet into a global stratospheric loading estimate. The average multiplier for deposited material in central Greenland varies from 1.2 x 10^9 for a high-latitude eruption to 2.4 x 10^9 for an equatorial eruption [*Clausen and Hammer*, 1988; p. 21].

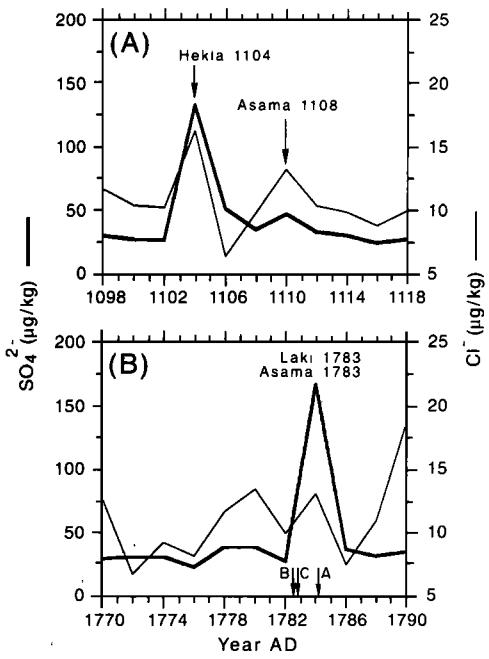

Figure 1. Time series of SO_4^{2-} and Cl^- concentrations from bi-yearly samples for twenty-year periods around the large historical Mt. Asama eruptions of (A) 1108 and (B) 1783. Age scale is mid-summer for year given. Age of samples containing volcanic glass shown in (B) by arrows A-C.

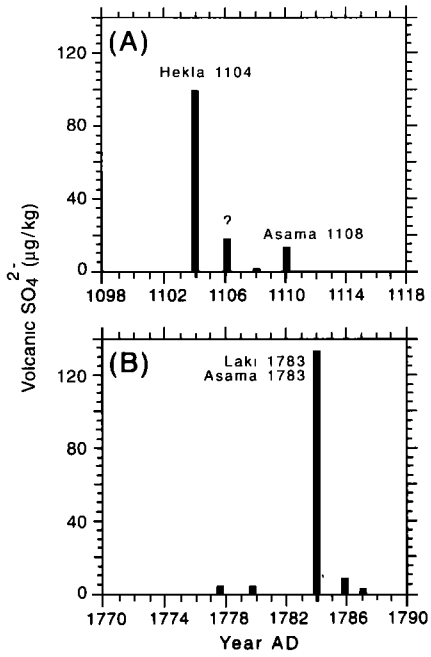

Figure 2. Volcanically-derived SO_4^{2-} associated with individual eruptions for twenty-year time periods around the (A) 1108 and (B) 1783 eruptions of Asama. SO_4^{2-} values from residuals above robust spline fit of original SO_4^{2-} time series as presented in *Zielinski et al.* [1994].

We used the mid-point of these two values (i.e., 1.8 x 10^9) to estimate stratospheric loading for the mid-latitude Asama volcano. We also are assuming that all transport to Greenland from Asama eruptions was stratospheric, which is probably a valid assumption given the distance between sites. The amount of volcanically-derived SO_4^{2-} used in our calculations is equal to the residual above a robust spline smoothing of the original SO_4^{2-} time series [*Zielinski et al.*, 1994].

Results

We were able to identify peaks in both the SO_4^{2-} and Cl^- time series for the 1108 Asama eruption, although the October 1104 Hekla (Iceland) eruption is much more prominent (Figure 1A). The presence of the large Hekla signal provides a reliable time-line [*Meese et al.*, 1994] that supports the age of layers in this section of core and thus the presence of Asama aerosols. No other VEI ≥ 4 eruptions are known to have occurred in the decade from 1100 to 1110 [*Simkin et al.*, 1981]. There appears to be almost a two-year lag between the eruption and SO_4^{2-} deposition in Greenland (i.e., in 1110) further suggesting stratospheric transport and validating the use of *Clausen and Hammer's* [1988] technique. A very large SO_4^{2-} peak during the summer of 1784 represents aerosols from the Laki and Asama events (Figure 1B).

The amount of volcanically-produced SO_4^{2-} associated with each Asama event is more clearly shown by the plot of SO_4^{2-} residuals (Figure 2) that we previously developed [*Zielinski et al.*, 1994]. Residual value for the 1108 event is 14 µg/kg (Table 1). Note that in our previous work we used a SO_4^{2-} residual value of 25 µg/kg as the minimum value for a significant climate-forcing eruption. The large magnitude of the 1783 signal (175 µg/kg) is clearly shown (Figure 2B).

Using these residual values we initially estimated a total stratospheric loading of H_2SO_4 for the 1108 event as we feel this signal may reflect deposition of only Asama aerosols

Table 1. Estimates of Stratospheric Mass Loading of H_2SO_4 and Optical Depth for A.D. 1108 and A.D. 1783 Eruptions of Mt. Asama from the GISP2 Ice Core

	1108	1783
Volcanic SO_4^{2-}(μg/kg)	14	4.7*
Volcanic SO_4^{2-} Flux (x10^3 kg/km^2)†	5.8	2.0
Stratospheric Loading H_2SO_4 (Tg)&	10.4	3.5
Total Stratospheric Loading (Tg) #	13	4.3
Stratospheric Optical Depth (τ_D)¶	0.087	0.029

*Based on assumption that sulfur production from the 1783 eruption is one-third that of the 1108 eruption.

†Product of volcanic SO_4^{2-}, length of ice and ice density

&Product of volcanic SO_4^{2-} flux and a multiplier of 1.8 x 10^9 for a mid-latitude eruption. See text for explanation.

#Composition, by weight, of H_2SO_4 aerosol particles is roughly 75% H_2SO_4 and 25% H_2O. *Self et al.* [in press] used 1.25 as an average multiplier to convert to total loading.

¶Determined from relationship defined by Stothers [1984], stratospheric mass loading (M_D) = 1.5 x 10^{14} τ_D g.

(Table 1). Our estimate for the 1108 event is an order of magnitude greater than that derived from the petrologic method [*Kohno et al.*, 1993]. *Rampino and Self* [1984] showed that stratospheric loading estimates from the petrologic method are often an order of magnitude less than that from optical depth and ice core calculations. Using our loading numbers we then calculated the potential stratospheric loading of the 1783 Asama eruption (Table 1) assuming that sulfur production from the 1108 eruption was three times greater based on petrologic analyses [*Kohno et al.*, 1993]. Finally, we suggest the potential stratospheric optical depth (τ_D) for each eruption given the mass loadings that we calculated in Table 1.

We found volcanic glass in several samples from the 1783 section of core including basaltic shards that are similar to that from the Laki eruption [Glass C, Figures 1B and 3; *Fiacco et al.*, in press]. A high-alkali andesitic to dacitic glass was found on two filters corresponding to layers from early in 1783 (Glass B; Figures 1B and 3) and late in 1784 (Glass A; Figures 1B and 3). The composition of these shards do not match glass from the plinian portion of the eruption and only a few shards from the pyroclastic flow component overlap with the composition of GISP2 particles (Figure 3). Glass with 62-64% silica has yet to be found in Laki deposits (K. Grönvold, personal communication). We have not yet sampled layers from 1110.

Discussion and Conclusions

Our estimations of H_2SO_4 loading for the 1783 and 1108 (Table 1) Asama eruptions range between 3 and 10 Tg with resulting τ_D values of 0.029 and 0.087, respectively. These values are low compared to direct measurements for more recent eruptions over the last couple of centuries known to have perturbed climate (e.g., 0.10-0.15 for El Chichón), although mass loading and τ_D for the 1108 eruption is similar to, or slightly greater than Northern Hemisphere values in the mid-1960s that may include the Agung eruption [*Rampino et al.*, 1988; *Sato et al.*, 1993]. Thus, sulfur loading from the 1783 eruption of Asama probably was not of sufficient magnitude (τ_D=0.029) to induce major climatic cooling in the Northern Hemisphere. Estimates of stratospheric optical depths for most climatically-forcing volcanism since A.D. 1850 are much greater than 0.02-0.03 [Figure 1 in *Sato et al.*, 1993]. Furthermore, the large signal around 1783 found in ice cores from Greenland is most likely almost entirely due to aerosols from Laki. Even if sulfur deposition for the 1108 and 1783 eruptions were essentially equal, a volcanic flux of 5.8 μg/kg (Table 1) is an order of magnitude less than the 57 μg/kg flux that we calculated from biyearly samples for the total 1783 signal. Estimates of the stratospheric loading of H_2SO_4 for the Laki eruption range from about 100 to 200 Tg [e.g., *Palais and Sigurdsson*, 1989; *Clausen and Hammer*, 1988]; estimates from Greenland ice cores are complicated because there is a tropospheric component contributing to the Laki signal. Nevertheless, we conclude that most volcanically-induced climatic and atmospheric phenomena during the mid-1780s were due to the Laki eruption.

One potential bias that we must address is any difference in the poleward transport of Asama aerosols due to season of the eruption (May 1108 versus August 1783) that could result in under-representation of 1108 aerosol deposition. However, we do not feel that is a problem based on previous atmospheric studies related to the May 1980 eruption of Mt. St. Helens. *McCormick and Trepte* [1987] detected an increase in stratospheric optical depth in the Arctic three months after the eruption suggesting very quick transport to polar regions from this mid-latitude eruption. Quick transport would limit the amount of loss, although there could still be some loss in the two years prior to deposition. The multiplier invoked by *Clausen and Hammer* [1988] should account for any aerosol loss in the polar regions prior to deposition. Interestingly, SO_4^{2-} levels in the mid-1780s return to background levels by early 1785 [*Fiacco et al.*, in press] meaning either quicker transport and deposition of aerosols from the 1783 Asama eruption or very little (if any) deposition in Greenland.

To further substantiate our suggestion that the 1783 Asama eruption had very little climatic impact, we evaluated the climatic effects of the 1108 eruption by compiling proxy data (e.g., tree-ring records) around the 1108 time period. If minimal evidence of climatic cooling exists in the Northern Hemisphere following the 1108 event then intuitively the 1783 eruption would have had essentially no hemispheric effects. In fact, this is just the case. The only suggestion of climatic cooling around 1108, particularly in Europe, is a reconstructed summer temperature minimum in 1109 from Fennoscandia tree rings, a value that is the second coldest over the 1500-year record [*Briffa et al.*, 1990]. However, a refinement of the technique [*Briffa et al.*, 1992] failed to place 1109 in the top five coolest summers over that same time span. Similarly, there are no records of climatic cooling following 1108 in German oak chronologies [*Lamb, 1977*; p. 595-602],

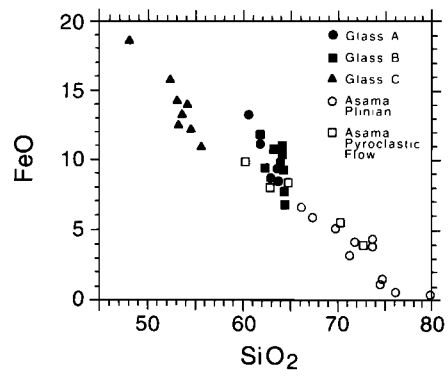

Figure 3. Covariation plot of volcanic glass found in the 1783/84 section of the GISP2 core as compared to the range in composition of glass shards from both plinian and pyroclastic flow deposits of the 1783 Asama eruption.

in Irish and Scottish oaks [*Baillie*, 1977], nor in Bristlecone pine records of America [*LaMarche and Hirschboek*, 1984].

We do not want to say, however, that Laki was the sole eruption responsible for climatic perturbations in the mid-1780s. Benjamin Franklin, in his pioneering work linking volcanic eruptions to climate over 200 years ago, indicated that several eruptions may have contributed to the dry fog observed in Europe during the summer of 1783 [quoted in *Sigurdsson*, 1982]. *Wood* [1992] noted the appearance of dry fog in Europe and of cooler and wetter conditions in Japan prior to either the Laki or Asama eruption. He also concluded that there may have been another eruption earlier in 1783 that contributed to atmospheric conditions at that time. The appearance of glass in the GISP2 core that is predominantly outside the known compositional range of Asama and Laki glass (Figure 3) could support the presence of another eruption.

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