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Key Points:
- The 1835 Cosigüina eruption released large amounts of sulfur to the atmosphere
- Sulfur was partitioned among sulfides, andesite melt, and a fluid phase
- Tree ring chronologies record global cooling in response to the 1835 eruption

Sulfur budget and global climate impact of the A.D. 1835 eruption of Cosigüina volcano, Nicaragua

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Abstract Large explosive volcanic eruptions can inject massive amounts of sulfuric gases into the Earth's atmosphere and, in so doing, affect global climate. The January 1835 eruption of Cosigüina volcano, Nicaragua, ranks among the Americas' largest and most explosive historical eruptions, but whether it had effects on global climate remains ambiguous. New petrologic analyses of the Cosigüina deposits reveal that the eruption released enough sulfur to explain a prominent circa A.D. 1835 sulfate anomaly in ice cores from both the Arctic and Antarctic. A compilation of temperature-sensitive tree ring chronologies indicates appreciable cooling of the Earth's surface in response to the eruption, consistent with instrumental temperature records. We conclude that this eruption represents one of the most important sulfur-producing events of the last few centuries and had a sizable climate impact rivaling that of the 1991 eruption of Mount Pinatubo.

1. Introduction

The main factor controlling the climatic impact of a given volcanic eruption is the total amount of sulfur that gets pumped into the stratosphere [*Robock*, 2000]. There, sulfur compounds are converted to H_2SO_4 aerosol particles that circulate around the globe and reflect sunlight back to space, resulting in net cooling at the surface. Such radiative forcing of climate has been directly observed for the twentieth century eruptions of Agung, Indonesia, in 1963, El Chichón, Mexico, in 1982, and Pinatubo, Philippines, in 1991 [*Robock*, 2000] and has been assumed for earlier historical eruptions (e.g., Huaynaputina, Peru, in 1600 [*de Silva and Zielinski*, 1998] and Tambora in 1815 and Krakatau in 1883, both in Indonesia [*Rampino and Self*, 1982; *Self et al.*, 2004]).

The January 1835 eruption of Cosigüina volcano (87.57°W, 12.98°N, 872 m altitude), in northwest Nicaragua, sent ash to Mexico, Costa Rica, and perhaps as far as Colombia and Jamaica, producing a bulk tephra volume of ~6 km^3 (volcanic explosivity index (VEI) = 5) [*Newhall and Self*, 1982; *Self et al.*, 1989; *Scott et al.*, 2006]. Initial estimates of eruptive volume in excess of >50 km^3 led a number of workers to assume that the eruption had an important climatic impact and was responsible for cool surface temperatures in the middle-to-late 1830s [*Lamb*, 1970; *Angell and Korshover*, 1985]. However, *Self et al.* [1989] showed that these volumes were much too large. Based on a revised eruptive volume, apparently low-sulfur yield [*Palais and Sigurdsson*, 1989] and ambiguous climate trends of the 1830s, *Self et al.* [1989] proposed a small to negligible atmospheric impact of the Cosigüina eruption. Yet subsequently, several studies, particularly of ice cores [e.g., *Gao et al.*, 2008] and tree rings [e.g., *Briffa et al.*, 1998], have listed the 1835 eruption as a possible climate-effective event. Here we use ice core data acquired since the late 1980s and new analyses of matrix glasses, melt, fluid, and sulfide inclusions from the 1835 deposits to reassess the sulfur budget of the Cosigüina eruption. We also utilize instrumental temperature records and compile data from temperature-sensitive tree ring chronologies to reevaluate the impact of the eruption on global climate.

2. A Sulfur-Producing Eruption

2.1. Ice Cores

It is well established that sulfate deposits in polar ice sheets record past explosive, sulfur-producing volcanic eruptions and can be used to estimate the original amount of sulfur that these eruptions injected into the stratosphere [*Zielinski*, 1995; *Gao et al.*, 2008]. Since the late 1980s, several studies of ice cores have identified volcanic sulfate horizons dated at circa A.D. 1835 [e.g., *Legrand and Delmas*, 1987; *Moore et al.*, 1991; *Delmas et al.*, 1992; *Langway et al.*, 1995; *Zielinski*, 1995; *Cole-Dai et al.*, 2000; *Budner and Cole-Dai*, 2003; *Crowley et al.*, 2008; *Ferris et al.*, 2011; *Plummer et al.*, 2012]. These signals are almost certainly due to the
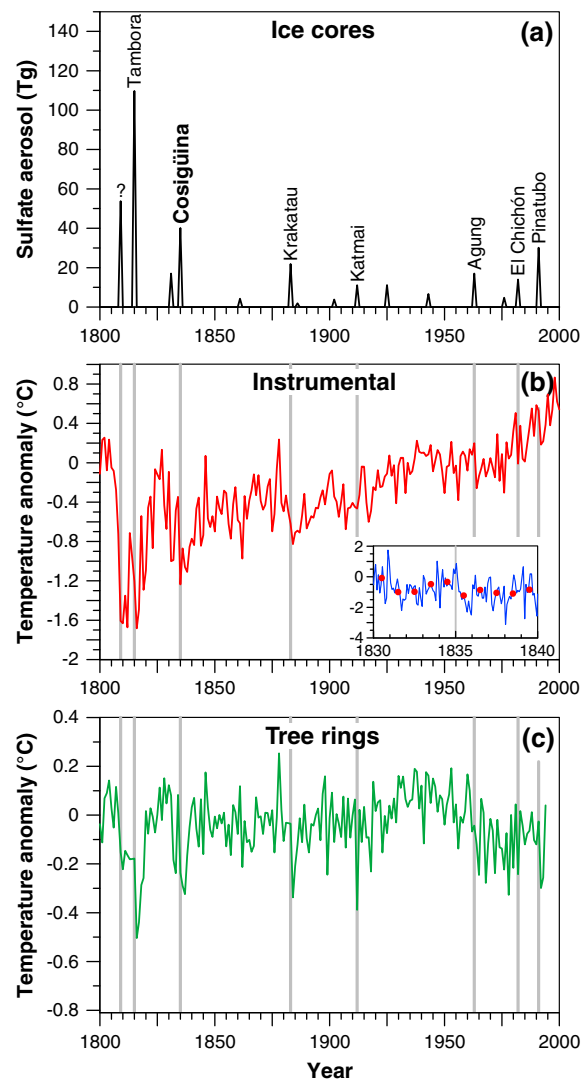
Figure 1. Ice core, instrumental, and tree ring evidence for a global impact of the 1835 Cosigüina eruption. (a) Annual global stratospheric volcanic sulfate aerosol injections since A.D. 1800 based on a compilation of data from 54 ice cores from both polar regions [*Gao et al.*, 2008]. (b) Instrumental temperature records compiled by Berkeley Earth (www.berkeleyearth.org) [*Rohde et al.*, 2013]. Annual summary of "All Land" average temperature is reported as anomalies relative to the 1951–1980 average. The inset shows the same data (red circles, plotted in the middle of the year) and detailed monthly averages (blue line) for the years 1830–1839. (c) Estimated Northern Hemisphere summer temperature anomalies relative to the 1881–1960 period based on tree ring density data [*Briffa et al.*, 1998]. For Figures 1b and 1c, grey bars correspond to the years of major sulfur-producing volcanic eruptions labeled in Figure 1a.
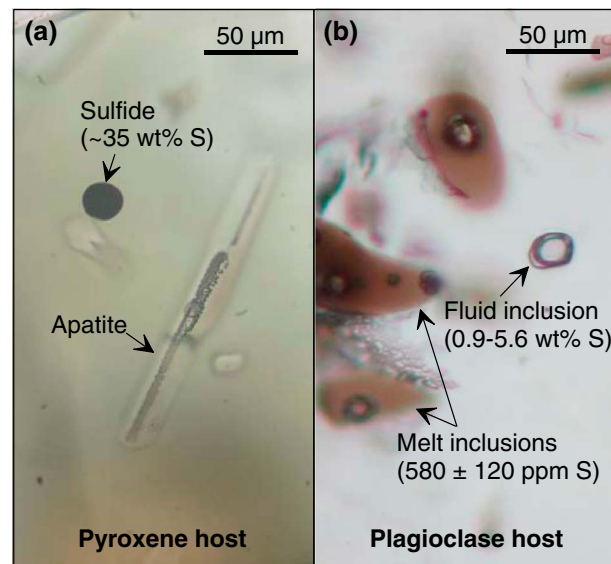
eruption of Cosigüina, which was 3 orders of magnitude larger than any other contemporaneous eruption [*Siebert and Simkin*, 2002]. A data compilation from 54 ice cores from across the Arctic and Antarctic provides the most comprehensive record available of sulfur-producing volcanic events in the past 1500 years [*Gao et al.*, 2008]. Since the beginning of the nineteenth century, the five most prominent sulfate spikes in ice cores correspond, in decreasing order of importance, to the eruptions of Tambora in 1815, an unidentified volcano in 1809, Cosigüina in 1835, Pinatubo in 1991, and Krakatau in 1883 (Figure 1a). A small peak present for the year 1831 may relate to the VEI = 4? eruption of Babuyan Claro, Philippines [*Siebert and Simkin*, 2002]. Notably, both Cosigüina (12.98°N) and Pinatubo (15.13°N), despite their Northern Hemisphere locations, produced significant sulfate deposition at the South Pole [*Cole-Dai and Mosley-Thompson*, 1999]. *Gao et al.* [2008] calculated a global stratospheric sulfate aerosol injection of 40.2 Tg (13.4 Tg S) for the 1835 Cosigüina eruption, compared to 30.1 Tg (10.0 Tg S) for the 1991 Pinatubo eruption. For Pinatubo, this estimate is consistent with values derived from petrologic [*Scaillet et al.*, 2003] and remote sensing methods [*Guo et al.*, 2004] (Table S1 in the supporting information).

2.2. Petrology
2.2.1. Preeruptive Magmatic Conditions
A detailed petrologic analysis of the 1835 eruptive products is presented by *Longpré et al.* [2014]; here we briefly summarize the main findings of this study. The 1835 eruption tapped a compositionally zoned magma reservoir; a subordinate volume of dacite and silicic andesite magma was erupted first, followed by andesite which represents the bulk of the erupted magma [*Scott et al.*, 2006]. The dacite and silicic andesite are related to the andesite through crystal-liquid segregation and fractional crystallization processes, and the 1835 magmas likely have a pre-1835 basaltic andesite parent. Mineral-melt thermometry indicates that temperatures in the reservoir ranged from 930°C for the dacite to 1010°C for the andesite, whereas oxygen

fugacity (fO_2) conditions extended from the nickel-nickel oxide (NNO) buffer to 1 log unit above it. Preeruptive dissolved water concentrations, estimated by the volatiles-by-difference method [e.g., *Humphreys et al.*, 2006] and plagioclase-melt hygrometry [*Lange et al.*, 2009], were likely between 3.2 and 4.2 wt %, suggesting a low-pressure (~100 MPa) magma reservoir at about 4 km depth. The occurrence of phenocryst-hosted sulfide globules and H_2O-rich fluid inclusions demonstrates that the 1835 magmas

Figure 2. Microscopic evidence for the sulfur budget of Cosigüina magmas. Photomicrographs of (a) a pyroxene-hosted sulfide globule and (b) plagioclase-hosted melt and fluid inclusions in 1835 andesite.

had reached saturation with both a sulfide phase and a hydrous fluid prior to eruption (Figure 2). Below, we first calculate the theoretical amount of sulfur available in the 1835 system, and then we consider the contributions of the sulfide droplets, andesite melt, and hydrous fluid to the sulfur budget of the eruption.

2.2.2. Theoretical Total Sulfur Available

Mafic magmas feeding Central American volcanoes commonly contain up to 2000–2500 ppm S [*Sadofsky et al.*, 2008]. For Cosigüina, a plot of sulfur versus FeO* in melt inclusions and matrix glasses shows an overall positive correlation (as discussed in the following section, this is attributed to the formation of sulfides) and highest S values of 1000–1200 ppm for pre-1835 basaltic andesite melt inclusions (Figure 3b, see also Figure S1). However, the basaltic andesite melt inclusions, which are

hosted in Fo_{71-74} olivine, display a wide range of S at similar FeO*; this range is likely due to degassing, and we argue that mafic parental Cosigüina magmas contain at least 1000–1200 ppm S. Taking 1200 ppm as initial concentration and using enrichment factors of incompatible elements, such as K and Rb, from basaltic andesite to dacite (Figures 3d and S2), we calculate that S, had it behaved incompatibly, would have reached a concentration of ~2100 ppm in andesite and ~2700 ppm in dacite. Assuming a closed system, this theoretical concentration (C_T), when scaled to the magma mass (m_m) erupted in 1835, gives a rough approximation of the total sulfur (S_T in Tg of S) that may have been available in the Cosigüina reservoir:

$$S_T = \left(1 \times 10^{-15}\right) m_m C_T$$

and

$$m_m = \rho_m \phi_m V$$

where the constant 1×10^{-15} is a conversion factor for sulfur concentration (ppm/kg) into sulfur mass (Tg), ρ_m is the density of the andesite magma (2500 kg/m^3), ϕ_m is the melt fraction (0.8, based on the average crystallinity of the andesite), and V is the dense rock equivalent magma volume emitted (2.4–3.7 km^3, based on a bulk deposit volume of 6 km^3 and bulk deposit density of 1000–1550 kg/m^3 [*Scott et al.*, 2006]). These equations yield a magma mass of 4800–7400 Tg, and theoretical total sulfur mass of 10–15 Tg. Although important, this calculation does not capture the complexity of the sulfur budget, which likely involved loss and/or gain of S by open system processes and partitioning of S into multiple phases [*Wallace and Edmonds*, 2011]. It is nevertheless interesting to note the excellent agreement of this theoretical sulfur amount with the ice core estimate of ~13 Tg.

2.2.3. Sulfur Sequestered by Sulfides

Rare and tiny (5–20 μm) sulfides occur as inclusions within phenocrysts in all 1835 samples (Figure 2a) but were not found in matrix glasses or in the pre-1835 basaltic andesite. Electron microprobe analysis of seven sulfides reveals the existence of two types: (1) more common Fe sulfide, i.e., pyrrhotite and (2) Fe-Cu sulfide, with a composition approaching that of cubanite, which we found in the andesite only (Table S2). As documented by *Lowenstern* [1993] for Katmai, these magmatic sulfides likely sequestered significant amounts of sulfur in the 1835 magma reservoir. This is consistent with the positive correlation of S and FeO* in melt inclusions, which reflects the sulfur concentration at sulfide saturation (SCSS) calculated after *Liu et al.* [2007] (Figure 3b). In addition, copper concentrations in bulk rocks decline precipitously from basaltic andesite to dacite (Figure S1). We estimate that removal of ~0.5 wt % sulfides during magma differentiation
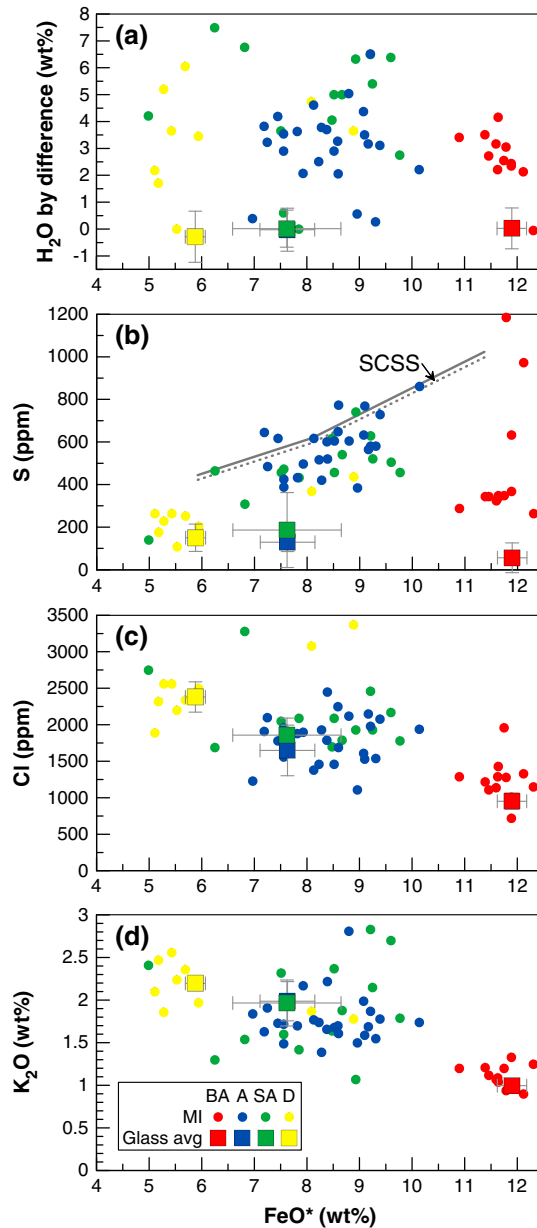
Figure 3. Composition of pre-1835 basaltic andesite (BA) and 1835 andesite (A), silicic andesite (SA) and dacite (D) melt inclusions (MI, circles), and matrix glasses (squares). For matrix glasses, error bars are 1 standard deviation of the mean (Table S3). (a) H_2O by difference, (b) S, (c) Cl, and (d) K_2O versus FeO*. In Figure 3b, the sulfur concentration at sulfide saturation (SCSS) calculated after *Liu et al.* [2007] is shown for $fO_2 = NNO$ (solid line) to NNO + 1 (dotted line).

from a pre-1835 basaltic andesite parent could potentially account for the difference between theoretical (~2100 ppm) and observed (380–860 ppm) S concentrations in andesite and the Cu depletion. Such fractionation would have provided a sink for 24–37 Tg of sulfide, or 8.4–13 Tg S. However, the apparent absence of sulfides in 1835 matrix glasses raises the question as to whether they were stable just prior to eruption; sulfide decomposition through interaction with a fluid phase may be common in magmatic systems [e.g., *Nadeau et al.*, 2010; *Berlo et al.*, 2014]. We propose that such a mechanism of sulfide breakdown may have remobilized significant amounts of S at Cosigüina.

2.2.4. Sulfur Dissolved in the Andesite Melt

Measured dissolved H_2O (by difference), S, and Cl contents in 1835 matrix glasses and plagioclase-hosted melt inclusions (Table S3) constrain syneruptive devolatilization of the magmatic liquid. The matrix glasses show lower H_2O, S, and Cl concentrations compared to melt inclusions for all three magma compositions, indicating substantial degassing (Figures 3 and S1). In particular, sulfur concentrations in melt inclusions from the andesite scoria range from 380 to 860 ppm, with an average of 580 ± 120 ppm (1σ). In contrast, the sulfur content of the andesite matrix glass is 130 ± 40 ppm. Our results differ markedly from those of *Palais and Sigurdsson* [1989], who reported indistinguishable matrix glass and melt inclusion sulfur concentrations of 360 ppm. Replicate analyses of the well-characterized VG-2 glass standard (1400 ppm S [*Jochum et al.*, 2005]) during our electron microprobe sessions yielded 1440 ± 60 ppm S ($n = 21$), placing a high level of confidence in our results.

We calculated the sulfur emission (S_E) by syneruptive devolatilization of the andesite melt using the following equation [e.g., *Devine et al.*, 1984; *Palais and Sigurdsson*, 1989]:

$$S_E = \left(1 \times 10^{-15}\right) m_m \Delta S_m$$

where ΔS_m is the sulfur lost from the andesite melt during the eruption (580 ± 120 ppm in melt inclusions—130 ± 40 ppm in matrix glass = 290–610 ppm degassed). This calculation indicates that the Cosigüina magma discharged at least 1.4–4.5 Tg of S to the atmosphere, along with 40–430 Tg of H_2O and 0–6.1 Tg of Cl. For sulfur, these figures for Cosigüina are much higher than the 0.1 Tg released by degassing of the Pinatubo dacite [*Westrich and Gerlach*, 1992], yet they are 3 to 10 times lower than estimates based on ice cores [*Gao et al.*, 2008] (Table S1). The classic "petrologic method" as used here generally underestimates actual sulfur emissions of large explosive eruptions, a discrepancy best explained by the presence of an exsolved fluid phase in the magma reservoir [*Wallace*, 2001].

2.2.5. Sulfur Exsolved in the Fluid Phase

The occurrence of phenocryst-hosted fluid inclusions in all 1835 samples provide direct evidence for the existence of a fluid phase within Cosigüina's magma reservoir prior to eruption (Figure 2b). Phase relationships determined by microthermometry indicate that this fluid contained about 87 wt % H_2O, 7 wt % CO_2, and 6 wt % NaCl equivalents and had a density of ~200 kg/m^3. For a temperature of ~1000°C, this implies a fluid entrapment pressure of ~100 MPa (see supporting information), in excellent agreement with the pressure estimate derived from H_2O contents in melt inclusions. Sulfur is known to partition strongly into the fluid phase [*Webster and Botcharnikov*, 2011, and references therein] and is thus also expected as a component of the fluid inclusions. At crustal pressures and temperatures, partitioning of sulfur between silicate melt and a coexisting fluid phase is a complex function of melt composition and oxygen fugacity in particular. For intermediate to silicic magmas and oxygen fugacity conditions of NNO to NNO + 1 (applicable to Cosigüina), recent experiments indicate S_{fluid}/S_{melt} values between 20 and 80 [*Keppler*, 2010; *Zajacz et al.*, 2012]. For Cosigüina, this translates into 0.9 to 5.6 wt % S in the fluid phase, assuming equilibrium with a melt with 580 ± 120 ppm S. Another approach is to use our salinity estimate and a molar Cl/S ratio of 1.7 in the fluid [*Shinohara*, 2013]; for this method, we obtain 2.4 wt % S.

Based on incompatible element enrichment systematics and dissolved H_2O estimates in Cosigüina basaltic andesite to dacite (Figure 3, see also Figures S1 and S2), the 1835 andesite appears to have suffered preeruptive loss of ~1.4 wt % H_2O (3.3 ± 1.7 wt % observed versus 4.7 ± 1.8 wt % predicted), giving 1.6 wt % total fluid. Thus, considering a conservative amount of exsolved fluid equivalent to 1–2 wt % [cf. *Wallace*, 2001] of the magma mass erupted in 1835 (4800–7400 Tg), we calculate its contribution to the sulfur yield to be in the range 0.4–8.3 Tg S.

In summary, we can integrate the different repositories for sulfur in the 1835 system as follows: (1) a sink of 8.4–13 Tg S stored in sulfide droplets, (2) 1.4–4.5 Tg S dissolved in the andesite melt, and (3) 0.4–8.3 Tg S exsolved in the hydrous fluid. Assuming a passive role of sulfur in sulfides, combining repositories (2) and (3) gives a total sulfur yield range of 1.8–13 Tg. The upper end of this range agrees well with the estimate of ~13 Tg S from ice core data. However, as noted in section 2.2.3, some of the sulfur sequestered by sulfides may have been remobilized through sulfide breakdown [*Nadeau et al.*, 2010] and released to the atmosphere in the gas phase. Adding this potential contribution, we obtain a maximum S yield of 10–26 Tg S (median 18 Tg S). Since the value from ice cores accounts only for the sulfur that reached the stratosphere and made it to the poles in 1835, this estimate probably is a minimum for the total sulfur yield. The actual amount of sulfur emitted by the eruption hence may have well exceeded 13 Tg, which could easily be accounted for by sulfide decomposition or by slightly higher amounts of fluid (3–5 wt % total, rather than 1–2 wt %) in the magma reservoir.

3. Effect on Global Climate

3.1. Instrumental Temperature Records

Many authors have documented the climatic impact of volcanic eruptions using instrumental temperature records [e.g., *Kelly and Sear*, 1984; *Angell and Korshover*, 1985; *Bradley*, 1988; *Jones et al.*, 2003]. For the Cosigüina eruption, *Angell and Korshover* [1985], *Chenoweth* [1998], and *Jones et al.* [2003] found significant cooling reaching 0.3 to ~2°C over different parts of the Northern Hemisphere and the Bahamas for the years 1835–1838. In Figure 1b, we show the instrumental temperature data compiled by Berkeley Earth (www.berkeleyearth.org) [*Rohde et al.*, 2013]. These data are consistent with volcanically induced cooling for all but one (Katmai) of the major sulfur-producing eruptions of the past two centuries. Except for the strongly negative anomalies associated with the 1809 Unknown and 1815 Tambora eruptions, the most important effect is observed for Cosigüina, with cooling apparently starting rapidly in 1835 (see the inset in Figure 1b showing monthly data) and cool temperatures persisting until 1838. However, we note that the data exhibit a cooling trend starting as early as 1828 (see also Figure S4), suggesting that the Cosigüina eruption may not be the sole cause of the post-1835 cooling [cf. *Self et al.*, 1989].

3.2. Tree Rings

Temperature-sensitive tree ring chronologies, widely used to reconstruct past climates, also capture signals of global cooling induced by large sulfur-producing volcanic eruptions [e.g., *LaMarche and Hirschboeck*, 1984; *Briffa et al.*, 1998; *D'Arrigo et al.*, 2013; *Esper et al.*, 2013]. Figure 1c shows the tree ring-based summer temperature

Table 1. Tree Ring Evidence for Global Cooling Due to the Cosigüina Eruption[a]

Phenomenon and Criteria	Period	Location	Anomalous Years (Rank) 1830–1834	Anomalous Years (Rank) 1835–1839
Frost rings[b]	560–1969	Western U.S.	1831	1837
Light rings[c]	1398–1982	Northern Québec	None	1836
Low-density rings, 5% lowest densities[d]	1600–1975	North America, Europe	None	1837 (20th)
Low-density rings, 5% coldest summers[e]	1400–1994	Northern Hemisphere	None	1836 (21st) and 1837 (15th)
Low-density rings, 5% lowest densities[f]	1720–1993	Alaska, northern Canada	None	1836 (1st)
Narrow ring widths[g]	1595–1997	Kola Peninsula, Russia	?	1837
Narrow ring widths, 5% coldest years[h]	713–1995	Northern Hemisphere	None	1837 (8th)
Ring width minima and frost rings[i]	Last 5000 years	Western U.S.	None	1836 and 1838
Narrow ring widths, 5% coldest years[j]	910–2011	Northern Québec	None	1835 (9th), 1836 (23rd), 1837 (17th), 1838 (5th), and 1839 (28th)

[a]Years showing anomalous tree ring growth phenomena, according to the listed criteria, are reported for the periods 1830–1834 and 1835–1839. When applicable, ranks within the 5% most extreme values are given in parentheses. We note that D'Arrigo et al. [2013] argued that volcanic signals may be most clearly and quantitatively assessed using the maximum latewood density parameter [e.g., Briffa et al., 1998].
[b]LaMarche and Hirschboeck [1984].
[c]Filion et al. [1986].
[d]For the merged North America-Europe data of Jones et al. [1995].
[e]Briffa et al. [1998].
[f]Rank using post-A.D. 1720 data of D'Arrigo and Jacoby [1999] (their Table 2). Years were classified based on their recurrence and ranks within the 20 lowest density anomalies for each of the 13 chronologies. The most anomalous year within a chronology was given 20 points, the second most anomalous was given 19 points, etc., and points for each year were added for the 13 chronologies. According to this ranking, 1836 is the most anomalous year since A.D. 1720.
[g]Gervais and MacDonald [2001].
[h]D'Arrigo et al. [2006].
[i]Salzer and Hughes [2007].
[j]Gennaretti et al. [2014].

reconstruction of Briffa et al. [1998] for the Northern Hemisphere from A.D. 1800 to 1994. According to this reconstruction, virtually all sulfur-producing eruptions of the last two centuries (Figure 1a) are correlated with a marked negative temperature anomaly. Of the 10 coolest summers since A.D. 1800, eight closely follow major eruptions, i.e., 1816, 1817, 1818, 1836, 1837, 1884, 1912, and 1992. The Cosigüina and Pinatubo eruptions are associated with anomalies of −0.32 and −0.30°C for 1837 and 1992, respectively [Briffa et al., 1998]. We note that it has been proposed that tree ring-based reconstructions of paleotemperatures may underestimate cooling induced by volcanic eruptions [Mann et al., 2012; but see D'Arrigo et al., 2013].

In the last decades several other dendroclimatological studies have reported anomalous tree rings indicating cold temperatures for the years 1836 and 1837 and pointed to the Cosigüina eruption as the cause [LaMarche and Hirschboeck, 1984; Filion et al., 1986; Jones et al., 1995; D'Arrigo and Jacoby, 1999; Gervais and MacDonald, 2001; Salzer and Hughes, 2007; Gennaretti et al., 2014]. Table 1 compiles the results of these studies, based on wood density, ring width, or the occurrence of frost damage, for the periods 1830–1834 and 1835–1839. Aside from a frost ring reported for A.D. 1831 [LaMarche and Hirschboeck, 1984], there is no evidence that the years 1830–1834 were particularly cold. By contrast, in the studies for which relative ranking of years was possible, either 1836 or 1837 (or both) are within the 5% coldest years over several centuries; Gennaretti et al. [2014] even found anomalous ring widths for all years in the 1835–1839 period. The tree ring data thus provide strong support for substantial Northern Hemisphere cooling in response to the eruption of Cosigüina in 1835. Furthermore, a tropical sea surface temperature reconstruction, based on coral, tree ring, and ice core proxies, documents a −0.60°C anomaly for 1837 (the 25th most negative value over the past 400 years), suggesting that post-Cosigüina cooling extended to low latitudes [D'Arrigo et al., 2009].

4. Discussion and Conclusions

The analysis of Self et al. [1989] clearly showed that early workers [e.g., Lamb, 1970] were misguided by the exaggerated initial estimates of eruptive volumes and had overestimated the atmospheric impact of the 1835 Cosigüina eruption. With few data from ice cores and tree rings available at the time and based on inaccurate sulfur concentrations in Cosigüina matrix glasses and melt inclusions [Palais and Sigurdsson, 1989], Self et al. [1989] argued that the eruption had little or no effect on global climate. However, the evidence

accumulated from ice cores since the late 1980s, coupled with our confirmation that the Cosigüina magma contained appreciable amounts of dissolved sulfur and was associated with a fluid phase enriched in sulfur, leaves little doubt that the 1835 eruption was a major sulfur-producing event on the scale of the 1991 Mount Pinatubo eruption or larger. Cooling of global surface temperatures by 0.5 to 0.6°C followed the Pinatubo eruption [e.g., *Parker et al.*, 1996; *Self et al.*, 1996], and the Cosigüina eruption produced a similar effect. Indeed, tree ring chronologies indicate anomalously low Northern Hemisphere temperatures for 2 to 3 years after the eruption, consistent with instrumental temperature records. A recent study linking volcanological data for the Central American Volcanic Arc and global climate models is also consistent with post-Cosigüina cooling of 0.4°C [*Metzner et al.*, 2012]. The small sulfate spike in ice cores for A.D. 1831 (Figure 1a), the cooling trends from instrumental records (Figures 1b and S4), and the occurrence of frost rings in western U.S. pines for that year (Table 1) [*LaMarche and Hirschboeck*, 1984] may relate to the 1831 eruption of Babuyan Claro, in the Philippines. If so, the effects of this event may have enhanced the climatic impact of the Cosigüina eruption. Finally, as it likely delivered substantial Cl and Br to the stratosphere in addition to S, the 1835 eruption may have also caused significant ozone depletion [*Kutterolf et al.*, 2013].

References

Angell, J. K., and J. Korshover (1985), Surface temperature changes following the six major volcanic episodes between 1780 and 1980, *J. Clim. Appl. Meteorol.*, *24*, 937–951, doi:10.1175/1520-0450(1985)024%3C0937:STCFTS%3E2.0.CO;2.

Berlo, K., V. van Hinsberg, N. Vigouroux, J. Gagnon, and A. Williams-Jones (2014), Sulfide breakdown controls metal signature in volcanic gas at Kawah Ijen volcano, Indonesia, *Chem. Geol.*, *371*, 115–127, doi:10.1016/j.chemgeo.2014.02.009.

Bradley, R. S. (1988), The explosive volcanic eruption signal in Northern Hemisphere continental temperature records, *Clim. Change*, *12*, 221–243, doi:10.1007/BF00139431.

Briffa, K. R., P. D. Jones, F. H. Schweingruber, and T. J. Osborn (1998), Influence of volcanic eruptions on Northern Hemisphere summer temperature over the past 600 years, *Nature*, *393*, 450–455, doi:10.1038/30943.

Budner, D., and J. Cole-Dai (2003), The number and magnitude of large explosive volcanic eruptions between 904 and 1865 A.D.: Quantitative evidence from a new South Pole ice core, in *Volcanism and the Earth's Atmosphere*, *Geophys. Monogr. Ser.*, vol. 139, edited by A. Robock and C. Oppenheimer, pp. 165–176, AGU, Washington, D. C., doi:10.1029/139gm10.

Chenoweth, M. (1998), The early 19th century climate of the Bahamas and a comparison with 20th century averages, *Clim. Change*, *40*, 577–603, doi:10.1023/A:1005371320672.

Cole-Dai, J., and E. Mosley-Thompson (1999), The Pinatubo eruption in South Pole snow and its potential value to ice-core paleovolcanic records, *Ann. Glaciol.*, *29*, 99–105, doi:10.3189/172756499781821319.

Cole-Dai, J., E. Mosley-Thompson, S. P. Wight, and L. G. Thompson (2000), A 4100-year record of explosive volcanism from an East Antarctica ice core, *J. Geophys. Res.*, *105*, 24,431–24,441, doi:10.1029/2000JD900254.

Crowley, T. J., G. Zielinski, B. Vinther, R. Udisti, K. Kreutz, J. Cole-Dai, and E. Castellano (2008), Volcanism and the Little Ice Age, *PAGES News*, *16*, 22–23.

D'Arrigo, R., R. Wilson, and G. Jacoby (2006), On the long-term context for late twentieth century warming, *J. Geophys. Res.*, *111*, D03103, doi:10.1029/2005JD006352.

D'Arrigo, R., R. Wilson, and A. Tudhope (2009), The impact of volcanic forcing on tropical temperatures during the past four centuries, *Nat. Geosci.*, *2*, 51–56, doi:10.1038/ngeo393.

D'Arrigo, R., R. Wilson, and K. J. Anchukaitis (2013), Volcanic cooling signal in tree ring temperature records for the past millennium, *J. Geophys. Res. Atmos.*, *118*, 9000–9010, doi:10.1002/jgrd.50692.

D'Arrigo, R. D., and G. C. Jacoby (1999), Northern North American tree-ring evidence for regional temperature changes after major volcanic events, *Clim. Change*, *41*, 1–15, doi:10.1023/a:1005370210796.

de Silva, S. L., and G. A. Zielinski (1998), Global influence of the AD 1600 eruption of Huaynaputina, Peru, *Nature*, *393*, 455–458, doi:10.1038/30948.

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, doi:10.1034/j.1600-0889.1992.00011.x.

Devine, J. D., H. Sigurdsson, A. N. Davis, and S. Self (1984), Estimates of sulfur and chlorine yield to the atmosphere from volcanic eruptions and potential climatic effects, *J. Geophys. Res.*, *89*, 6309–6325, doi:10.1029/JB089iB07p06309.

Esper, J., L. Schneider, P. Krusic, J. Luterbacher, U. Büntgen, M. Timonen, F. Sirocko, and E. Zorita (2013), European summer temperature response to annually dated volcanic eruptions over the past nine centuries, *Bull. Volcanol.*, *75*, 1–14, doi:10.1007/s00445-013-0736-z.

Ferris, D. G., J. Cole-Dai, A. R. Reyes, and D. M. Budner (2011), South Pole ice core record of explosive volcanic eruptions in the first and second millennia A.D. and evidence of a large eruption in the tropics around 535 A.D, *J. Geophys. Res.*, *116*, D17308, doi:10.1029/2011JD015916.

Filion, L., S. Payette, L. Gauthier, and Y. Boutin (1986), Light rings in subarctic conifers as a dendrochronological tool, *Quat. Res.*, *26*, 272–279, doi:10.1016/0033-5894(86)90111-0.

Gao, C., A. Robock, and C. Ammann (2008), Volcanic forcing of climate over the past 1500 years: An improved ice core-based index for climate models, *J. Geophys. Res.*, *113*, D23111, doi:10.1029/2008JD010239.

Gennaretti, F., D. Arseneault, A. Nicault, L. Perreault, and Y. Bégin (2014), Volcano-induced regime shifts in millennial tree-ring chronologies from northeastern North America, *Proc. Natl. Acad. Sci. U.S.A.*, *111*, 10,077–10,082, doi:10.1073/pnas.1324220111.

Gervais, B. R., and G. M. MacDonald (2001), Tree-ring and summer-temperature response to volcanic aerosol forcing at the northern tree-line, Kola Peninsula, Russia, *Holocene*, *11*, 499–505, doi:10.1191/095968301678302940.

Guo, S., G. J. S. Bluth, W. I. Rose, I. M. Watson, and A. J. Prata (2004), Re-evaluation of SO_2 release of the 15 June 1991 Pinatubo eruption using ultraviolet and infrared satellite sensors, *Geochem., Geophys., Geosyst.*, *5*, Q04001, doi:10.1029/2003GC000654.

Humphreys, M. C. S., S. L. Kearns, and J. D. Blundy (2006), SIMS investigation of electron-beam damage to hydrous, rhyolitic glasses: Implications for melt inclusion analysis, *Am. Mineral.*, *91*, 667–679, doi:10.2138/am.2006.1936.

Jochum, K. P., U. Nohl, K. Herwig, E. Lammel, B. Stoll, and A. W. Hofmann (2005), GeoReM: A new geochemical database for reference materials and isotopic standards, *Geostand. Geoanal. Res.*, *29*, 333–338, doi:10.1111/j.1751-908X.2005.tb00904.x.

Jones, P. D., K. R. Briffa, and F. H. Schweingruber (1995), Tree-ring evidence of the widespread effects of explosive volcanic eruptions, *Geophys. Res. Lett.*, *22*, 1333–1336, doi:10.1029/94GL03113.

Jones, P. D., A. Moberg, T. J. Osborn, and K. R. Briffa (2003), Surface climate responses to explosive volcanic eruptions seen in long European temperature records and mid-to-high latitude tree-ring density around the Northern Hemisphere, in *Volcanism and the Earth's Atmosphere*, *Geophys. Monogr. Ser.*, vol. 139, edited by A. Robock and C. Oppenheimer, pp. 239–254, AGU, Washington, D. C., doi:10.1029/139GM15.

Kelly, P. M., and C. B. Sear (1984), Climatic impact of explosive volcanic eruptions, *Nature*, *311*, 740–743, doi:10.1038/311740a0.

Keppler, H. (2010), The distribution of sulfur between haplogranitic melts and aqueous fluids, *Geochim. Cosmochim. Acta*, *74*, 645–660, doi:10.1016/j.gca.2009.10.010.

Kutterolf, S., T. H. Hansteen, K. Appel, A. Freundt, K. Krüger, W. Pérez, and H. Wehrmann (2013), Combined bromine and chlorine release from large explosive volcanic eruptions: A threat to stratospheric ozone?, *Geology*, *41*, 707–710, doi:10.1130/g34044.1.

LaMarche, V. C., and K. K. Hirschboeck (1984), Frost rings in trees as records of major volcanic eruptions, *Nature*, *307*, 121–126, doi:10.1038/307121a0.

Lamb, H. H. (1970), Volcanic dust in the atmosphere; with a chronology and assessment of its meteorological significance, *Philos. Trans. R. Soc. London*, *266*, 425–533, doi:10.1098/rsta.1970.0010.

Lange, R. A., H. M. Frey, and J. Hector (2009), A thermodynamic model for the plagioclase-liquid hygrometer/thermometer, *Am. Mineral.*, *94*, 494–506, doi:10.2138/am.2009.3011.

Langway, C. C., Jr., K. Osada, H. B. Clausen, C. U. Hammer, and H. Shoji (1995), A 10-century comparison of prominent bipolar volcanic events in ice cores, *J. Geophys. Res.*, *100*, 16,241–16,247, doi:10.1029/95JD01175.

Legrand, M., and R. J. Delmas (1987), A 220-year continuous record of volcanic H_2SO_4 in the Antarctic ice sheet, *Nature*, *327*, 671–676.

Liu, Y., N.-T. Samaha, and D. R. Baker (2007), Sulfur concentration at sulfide saturation (SCSS) in magmatic silicate melts, *Geochim. Cosmochim. Acta*, *71*, 1783–1799, doi:10.1016/j.gca.2007.01.004.

Longpré, M.-A., J. Stix, F. Costa, E. Espinoza, and A. Muñoz (2014), Magmatic processes and associated timescales leading to the January 1835 eruption of Cosigüina volcano, Nicaragua, *J. Petrol.*, *55*, 1173–1201, doi:10.1093/petrology/egu022.

Lowenstern, J. B. (1993), Evidence for a copper-bearing fluid in magma erupted at the Valley of Ten Thousand Smokes, Alaska, *Contrib. Mineral. Petrol.*, *114*, 409–421, doi:10.1007/BF01046542.

Mann, M. E., J. D. Fuentes, and S. Rutherford (2012), Underestimation of volcanic cooling in tree-ring-based reconstructions of hemispheric temperatures, *Nat. Geosci.*, doi:10.1038/ngeo1394.

Metzner, D., S. Kutterolf, M. Toohey, C. Timmreck, U. Niemeier, A. Freundt, and K. Krüger (2012), Radiative forcing and climate impact resulting from SO_2 injections based on a 200,000-year record of Plinian eruptions along the Central American Volcanic Arc, *Int. J. Earth Sci.*, 1–17, doi:10.1007/s00531-012-0814-z.

Moore, J. C., H. Narita, and N. Maeno (1991), A continuous 770-year record of volcanic activity from East Antarctica, *J. Geophys. Res.*, *96*, 17,353–17,359, doi:10.1029/91JD01283.

Nadeau, O., A. E. Williams-Jones, and J. Stix (2010), Sulphide magma as a source of metals in arc-related magmatic hydrothermal ore fluids, *Nat. Geosci.*, *3*, 501–505, doi:10.1038/ngeo899.

Newhall, C. G., and S. Self (1982), The volcanic explosivity index (VEI): An estimate of explosive magnitude for historical volcanism, *J. Geophys. Res.*, *87*, 1231–1238, doi:10.1029/JC087iC02p01231.

Palais, J. M., and H. Sigurdsson (1989), Petrologic evidence of volatile emissions from major historic and pre-historic volcanic eruptions, in *Understanding Climate Change*, *Geophys. Monogr. Ser.*, vol. 52, edited by A. Berger, R. E. Dickinson, and J. W. Kidson, pp. 31–53, AGU, Washington, D. C., doi:10.1029/GM052p0031.

Parker, D. E., H. Wilson, P. D. Jones, J. R. Christy, and C. K. Folland (1996), The impact of Mount Pinatubo on world-wide temperatures, *Int. J. Climatol.*, *16*, 487–497, doi:10.1002/(sici)1097-0088(199605)16:5<487:aid-joc39>3.0.co;2-j.

Plummer, C. T., M. A. J. Curran, T. D. van Ommen, S. O. Rasmussen, A. D. Moy, T. R. Vance, H. B. Clausen, B. M. Vinther, and P. A. Mayewski (2012), An independently dated 2000-yr volcanic record from Law Dome, East Antarctica, including a new perspective on the dating of the 1450s CE eruption of Kuwae, Vanuatu, *Clim. Past*, *8*, 1929–1940, doi:10.5194/cp-8-1929-2012.

Rampino, M. R., and S. Self (1982), Historic eruptions of Tambora (1815), Krakatau (1883), and Agung (1963), their stratospheric aerosols, and climatic impact, *Quat. Res.*, *18*, 127–143, doi:10.1016/0033-5894(82)90065-5.

Robock, A. (2000), Volcanic eruptions and climate, *Rev. Geophys.*, *38*, 191–219, doi:10.1029/1998RG000054.

Rohde, R., R. Muller, R. Jacobsen, S. Perlmutter, A. Rosenfeld, J. Wurtele, J. Curry, C. Wickham, and S. Mosher (2013), Berkeley earth temperature averaging process, *Geoinfor. Geostat. An Overview*, *1*, 1–13, doi:10.4172/2327-4581.1000103.

Sadofsky, S. J., M. Portnyagin, K. Hoernle, and P. van den Bogaard (2008), Subduction cycling of volatiles and trace elements through the Central American volcanic arc: Evidence from melt inclusions, *Contrib. Mineral. Petrol.*, *155*, 433–456, doi:10.1007/s00410-007-0251-3.

Salzer, M. W., and M. K. Hughes (2007), Bristlecone pine tree rings and volcanic eruptions over the last 5000 yr, *Quat. Res.*, *67*, 57–68, doi:10.1016/j.yqres.2006.07.004.

Scaillet, B., J. F. Luhr, and M. R. Carroll (2003), Petrological and volcanological constraints on volcanic sulfur emissions to the atmosphere, in *Volcanism and the Earth's Atmosphere*, *Geophys. Monogr. Ser.*, vol. 139, edited by A. Robock and C. Oppenheimer, pp. 11–40, AGU, Washington, D. C., doi:10.1029/139gm02.

Scott, W., C. Gardner, G. Devoli, and A. Alvarez (2006), The A.D. 1835 eruption of Volcán Cosigüina, Nicaragua: A guide for assessing local volcanic hazards, in *Volcanic Hazards in Central America*, edited by W. I. Rose et al., *412*, 167–187, Spec. Pap. Geol. Soc. Am., Boulder, Colo., doi:10.1130/2006.2412(09).

Self, S., M. Rampino, and M. Carr (1989), A reappraisal of the 1835 eruption of Cosigüina and its atmospheric impact, *Bull. Volcanol.*, *52*, 57–65, doi:10.1007/BF00641387.

Self, S., J.-X. Zhao, R. E. Holasek, R. C. Torres, and A. J. King (1996), The atmospheric impact of the 1991 Mount Pinatubo eruption, in *Fire and Mud: Eruptions and Lahars of Mount Pinatubo, Philippines*, edited by C. G. Newhall and R. S. Punongbayan, pp. 1089–1115, Univ. of Washington Press, Seattle, Wash.

Self, S., R. Gertisser, T. Thordarson, M. R. Rampino, and J. A. Wolff (2004), Magma volume, volatile emissions, and stratospheric aerosols from the 1815 eruption of Tambora, *Geophys. Res. Lett.*, *31*, L20608, doi:10.1029/2004GL020925.

Shinohara, H. (2013), Volatile flux from subduction zone volcanoes: Insights from a detailed evaluation of the fluxes from volcanoes in Japan, *J. Volcanol. Geotherm. Res.*, *268*, 46–63, doi:10.1016/j.jvolgeores.2013.10.007.

Siebert, L., and T. Simkin (2002), Volcanoes of the world: An illustrated catalog of Holocene volcanoes and their eruptions, *Global Volcanism Program Digital Information Series, GVP-3*, Smithsonian Institution, Washington, D. C. [Available at http://www.volcano.si.edu.]

Wallace, P. J. (2001), Volcanic SO_2 emissions and the abundance and distribution of exsolved gas in magma bodies, *J. Volcanol. Geotherm. Res.*, *108*, 85–106, doi:10.1016/s0377-0273(00)00279-1.

Wallace, P. J., and M. Edmonds (2011), The sulfur budget in magmas: Evidence from melt inclusions, submarine glasses, and volcanic gas emissions, *Rev. Mineral. Geochem.*, *73*, 215–246, doi:10.2138/rmg.2011.73.8.

Webster, J. D., and R. E. Botcharnikov (2011), Distribution of sulfur between melt and fluid in S-O-H-C-Cl-bearing magmatic systems at shallow crustal pressures and temperatures, *Rev. Mineral. Geochem.*, *73*, 247–283, doi:10.2138/rmg.2011.73.9.

Westrich, H. R., and T. M. Gerlach (1992), Magmatic gas source for the stratospheric SO_2 cloud from the June 15,1991, eruption of Mount Pinatubo, *Geology*, *20*, 867–870, doi:10.1130/0091-7613(1992)020<0867:MGSFTS>2.3.CO;2.

Zajacz, Z., P. A. Candela, P. M. Piccoli, and C. Sanchez-Valle (2012), The partitioning of sulfur and chlorine between andesite melts and magmatic volatiles and the exchange coefficients of major cations, *Geochim. Cosmochim. Acta*, *89*, 81–101, doi:10.1016/j.gca.2012.04.039.

Zielinski, G. A. (1995), Stratospheric loading and optical depth estimates of explosive volcanism over the last 2100 years derived from the Greenland Ice Sheet Project 2 ice core, *J. Geophys. Res.*, *100*, 20,937–20,955, doi:10.1029/95JD01751.