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Past 220 year bipolar volcanic signals: remarks on common features of their source volcanic eruptions

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ABSTRACT. During the past 220 years, prominent signals of non-sea salt sulfate ion ($nssSO_4^{2-}$) concentration exceeding the background level, including both marine biogenic and anthropogenic SO_4^{2-}, were found in shallow ice cores from site H15 in East Antarctica and Site-J in southern Greenland. They were mostly correlated with past explosive volcanic eruptions. On the basis of this result and published results of shallow ice cores and snow pits at various locations on the Antarctic and Greenland ice sheets, eight common signals were found, of which six were assigned to the following explosive eruptions: El Chichón, Mexico, in 1982; Agung, Indonesia, in 1963; Santa Maria, Guatemala, in 1902; Krakatau, Indonesia, in 1883; Cosiguina, Nicaragua, in 1835; an unknown volcano between 1831 and 1834; Tambora, Indonesia, in 1815; and an unknown volcano in 1809. Volcanic eruptions which have a potential to imprint their signals in both the Antarctic and Greenland ice sheets were characterized by (1) location in low latitudes between 20° N and 10° S, and (2) eruption column height ≥25 km, corresponding to a volcanic explosivity index (VEI) ≥5.

INTRODUCTION

Ice cores extracted from ice sheets and glaciers preserve records of past volcanism in local and remote areas (e.g. Hammer and others, 1980). An explosive volcanic eruption produces an eruptive column, which reaches altitudes higher than 10 km, and injects huge amounts of volcanic ash and gas (mainly H_2O, CO_2, SO_2, H_2S, HCl and HF) into the stratosphere. Of the volcanic materials, sulfur gases (SO_2 and H_2S) are oxidized to form sulfuric acid (H_2SO_4) aerosols in the stratosphere. These aerosols are suspended in the stratosphere for a few years after eruptions and are spread globally by atmospheric circulation. Reaching polar regions, they are eventually deposited onto the ice-sheet surface by both dry-fallout and washout processes.

Marine biogenic and anthropogenic H_2SO_4 are also deposited on the ice sheet. The marine biogenic non-sea salt sulfate ($nssSO_4^{2-}$) concentration of snowfall seasonally changes due to high production of the biogenic H_2SO_4 in summer (e.g. Osada, 1994). The anthropogenic $nssSO_4^{2-}$ concentration in the Arctic ice cores increases from the mid-19th century onwards due to an increase in sulfur compounds produced by combustion of solid and liquid fossil fuels in industrialized regions (e.g. Mayewski and others, 1990). On the other hand, deposition of the anthropogenic H_2SO_4 is negligible in the Antarctic region because of the long distance from the sources in the Northern Hemisphere. Detection of the volcanic $nssSO_4^{2-}$ concentration peaks (i.e. volcanic signal) requires estimation of non-volcanic $nssSO_4^{2-}$ concentrations corresponding to the biogenic and anthropogenic $nssSO_4^{2-}$ concentration. The non-volcanic $nssSO_4^{2-}$ concentration of the Antarctic and Arctic ice cores is defined as average concentration plus twice the standard deviation (2σ), assuming that the concentration exhibits a Gaussian distribution (Langway and others, 1994; Cole-Dai and others, 1997a; Robertson and others, 2001).

Up to 100 explosive eruptions with eruptive columns reaching the stratosphere have been recorded during the past 220 years (Simkin and Siebert, 1994). However, less than 20 significant volcanic signals from the past 220 years have been found in Antarctic and Arctic ice cores, respectively (e.g. Hammer and others, 1980; Legrand and Delmas, 1987; Moore and others, 1991; Zielinski, 1995; Cole-Dai and others, 1997a). This suggests that the volcanic aerosols derived from explosive eruptions are not always found as prominent volcanic signals in the polar ice cores.

In this study, we discuss an appropriate method to identify volcanic signals, and accordingly we detect volcanic signals for the past 220 years in shallow ice cores drilled at Site-J (66°52′ N, 46°16′ W; 2030 m a.s.l.), Greenland, and site H15 (69°04′ S, 40°47′ E; 1050 m a.s.l.), Antarctica (Fig. 1). We have suggested volcanic eruptions as sources of the volcanic signals found in the H15 and Site-J ice cores. Then we identify signals of common source eruptions found in the H15, Site-J and various ice cores and snow samples from the Arctic and Antarctic ice sheets (Fig. 1).

SAMPLES AND ANALYTICAL PROCEDURE

A 120.2 m deep ice core was drilled at site H15, East Antarctica, in 1991 (Fujii and others, 1995; Fig. 1). Mean annual air temperature at H15 is −21 °C, and annual snow-accumulation rate is 32 cm a^{-1} w.e. (Kohno and others, 1996). Electrical conductivity measurements (ECM), a measure of acidity, were immediately performed at Syowa station after recovery. The core was cut into 7–10 cm thick slices corresponding to a period of 2–4 months, in a low-temperature room (−20°C) at the National Institute of Polar Research, Tokyo (Kohno

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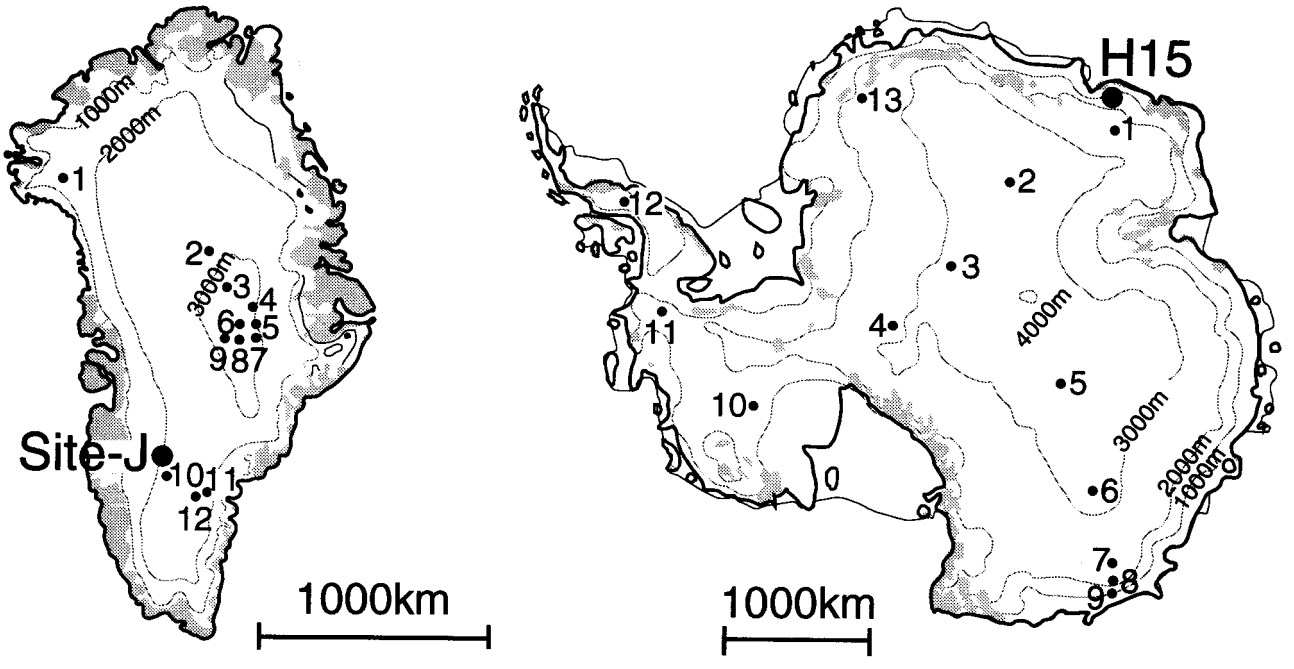

Fig.1. Location of Site-J ice-core drilling site in Greenland (left) and H15 ice-core drilling site in Antarctica (right). Other drilling sites mentioned in the text are numbered as follows. Greenland: 1. Camp Century; 2. North Central; 3. Summit, Site T; 4. Site E; 5. Site G; 6. Site A; 7. Crête; 8. Site B; 9. Site D; 10. Dye 2; 11. Dye 3; 12.20D. Antarctica: 1. G15; 2. Dome Fuji; 3. Plateau Remote; 4. South Pole; 5. Vostok; 6. Dome C; 7. D80; 8. D57; 9. D55; 10. Byrd Station; 11. Siple; 12. Dyer; 13. Amundsenisen.

and others, 1999). Surface 5–10 mm of the sliced samples were removed with a clean knife in order to eliminate contaminants. Water samples were used for the concentration measurements of major ions (Cl^-, NO_3^-, SO_4^{2-}, Na^+, K^+, Mg^{2+} and Ca^{2+}), pH and oxygen isotopic ratio ($\delta^{18}O$).

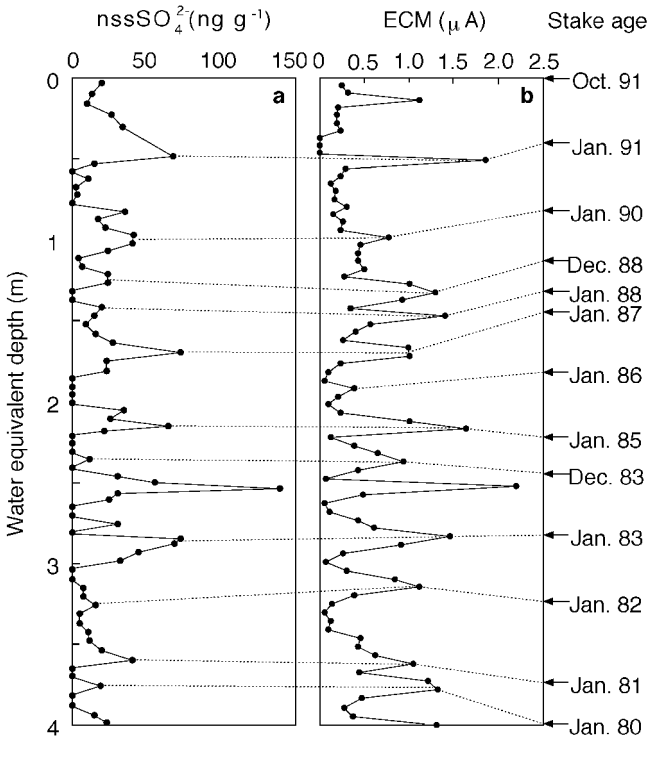

Fig. 2. (a) A profile of nssSO_4^{2-} concentration in firn from H15 as a function of water equivalent depth. A volcanic signal of El Chichón in 1982 is found at 2.54 m w.e. depth. (b) A profile of ECM in firn from H15. Stake ages (compiled in Kohno and others, 1996) are given to the right of the diagram (after Kohno and others, 1999, fig. 2). A nssSO_4^{2-} high tends to correspond to a ECM high.

A 205.15 m deep ice core was obtained at Site-J, Greenland, in 1989 (Watanabe and Fujii, 1990; Fig. 1). Mean annual air temperature at Site-J is −20°C, and annual snow accumulation rate is 42 cm a^{-1} ice equivalent (Shoji and others, 1991). The core was cut into 10 cm thick slices, which cover a period of 3 months, in a trench at the drilling site. The sliced samples were melted in Teflon containers in a microwave oven, after removal of the outer 5 mm part with a clean knife. Melted samples were used for in situ measurement of the electrical conductivity (EC) and for laboratory analyses of the $\delta^{18}O$ and micro-particle concentration. The water samples having high EC value were used for measurement of major-chemical composition and pH. Pooled samples for measurements of major chemical composition, pH and $\delta^{18}O$ were made from 10 successive samples, i.e. 100 cm long samples. Consequently, each pooled sample covers a period of 2 years.

Major-chemical compositions were measured by ion chromatography using a Dionex ion chromatograph (model 2000i) equipped with HIPC-AG4A and CG3 concentrators, AS4 and CS3 separator columns and anion and cation micro-membrane suppressors. Replicate measurements of standard solutions prepared from analytical grade reagents (NaCl, $NaNO_3$, K_2SO_4 and KCl, Mg^{2+} and Ca^{2+} 1000 ppm standard solutions) indicate that the relative analytical error is within 5% of the measured value. In this work, we only mention SO_4^{2-} and Na^+ concentrations in order to obtain nssSO_4^{2-} concentration, which is calculated by subtracting the SO_4^{2-} contribution of sea water from the total SO_4^{2-} concentration, assuming that Na^+ originates entirely from sea salt:

$$(SO_4^{2-})_{nss} = (SO_4^{2-})_{meas.} - (SO_4^{2-}/Na^+)_{seawater}(Na^+)_{meas.}, \quad (1)$$

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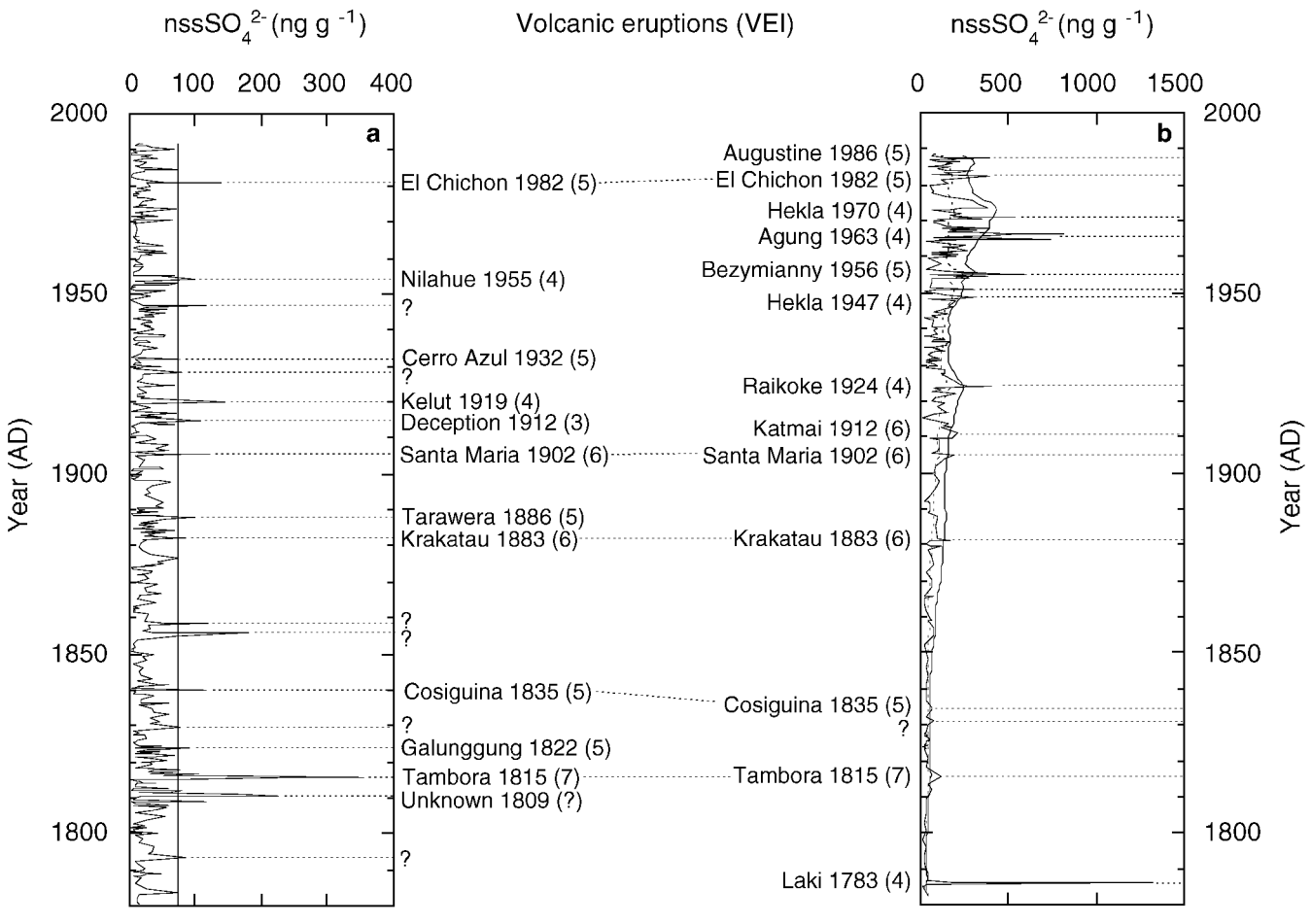

Fig. 3. (a) A profile of $nssSO_4^{2-}$ concentration against water equivalent depth of the H15 core (after Kohno and others, 1999, fig. 5). Thick solid line indicates a background concentration of $nssSO_4^{2-}$ (72 ng g^{-1}). Signals of $nssSO_4^{2-}$ concentration exceeding this background level are shown by dotted line, and the probable source eruptions with VEIs in parentheses are given to the right side of the diagram. (b) A profile of $nssSO_4^{2-}$ concentration against water equivalent depth of the Site-J core. Dashed and solid curves indicate the regression curve of $nssSO_4^{2-}$ concentration and a background level, respectively (see text). Signals of $nssSO_4^{2-}$ concentration exceeding this background are shown by dotted line, and the probable source eruptions with VEIs in parentheses are given to the left side of the diagram. Common volcanic signals among the signals of the H15 and Site-J cores also indicated by dotted line between (a) and (b).

where $(SO_4^{2-})_{meas.}$ and $(Na^+)_{meas.}$ are the measured SO_4^{2-} and Na^+ concentrations of the ice-core samples. $(SO_4^{2-}/Na^+)_{seawater}$, the ratio of SO_4^{2-} to Na^+ concentrations of sea water, was 0.25 in units of weight.

Acidity, which is an index of acidic aerosols deposited in the snow, was measured by a pH meter (HM-60S, TOA Electronics Ltd.). The $\delta^{18}O$ of samples was measured with a mass spectrometer (MAT-252) using H_2O–CO_2 exchange technique, and given by the conventional $\delta^{18}O$ expression with respect to Standard Mean Ocean Water.

DATING OF ICE CORES

The H15 core is dated by two independent methods: an empirical model of firn densification proposed by Herron and Langway (1980), and counting the number of ECM peaks due to high acidity of snow mainly from annual cycles of atmospheric $nssSO_4^{2-}$ concentration (Fujii, 1983; Hammer and others, 1986; Steffensen, 1988).

The atmospheric $nssSO_4^{2-}$ concentration is governed by production of marine biogenic dimethyl sulfide (DMS) and oxidation of the DMS to H_2SO_4 via SO_2 (Ayers and others, 1991). The production rate of the DMS increases in summer, and it is oxidized within 1 day (Koga and Tanaka, 1996). The acidity of snow is consequently high in summer precipitation. The $nssSO_4^{2-}$ concentration and acidity of drifting snow collected at Mizuho station, at a distance of 220 km from site H15, are significantly increased in summer precipitation (Osada, 1994). Therefore, the counts of the ECM and/or $nssSO_4^{2-}$ concentration peaks are used for dating the H15 core. Figure 2 compares the ECM and $nssSO_4^{2-}$ concentration profiles of the surface 4 m deep layer of the H15 core with the dates of the firn given by stake observations continuously performed at least once a year between January 1980 and October 1991 by the Japanese Antarctic Research Expedition (JARE Data Reports, 1981–92). The ECM and $nssSO_4^{2-}$ concentration peaks are assigned to summer layers of appropriate years determined by the stake observations. These peaks were correlated to the summer layers in all but a few cases. An extremely high $nssSO_4^{2-}$ concentration peak at 2.5 m depth corresponding to the latter half of 1983 is likely to be related to the sulfuric acid aerosol derived from an explosive volcanic eruption of El Chichón, Mexico, in 1982.

Errors of dating by the counting may increase with depth because of errors due to counting of non-summer peaks such as the volcanic signal at 2.5 m depth, or erroneous counting of small summer peaks. For the past 220 years the ages estimated

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by the ECM counting differ by a maximum of 10 years from the ages calculated by Herron and Langway's (1980) densification model (Kohno and others, 1999).

The Site-J core is precisely dated by Fujii and others (2001) using the following two methods: counting annual cycles of $\delta^{18}O$ and detection of reference peaks of atmospheric bomb-produced tritium (^{3}H) content. Three significant peaks of ^{3}H in the depth range 15–23 m correspond to the key horizons of 1959, 1963 and 1964 as detected in other ice cores from Greenland (e.g. Koide and others, 1982). The ages of these three peaks determined by the counted annual $\delta^{18}O$ cycles agreed with the ages of the key horizons. Dating errors are estimated to be within 2 years in the deeper part of the core (Fujii and others, 2001).

VOLCANIC SULFATE ION SIGNALS

A profile of $nssSO_4^{2-}$ concentration of the H15 core is shown in Figure 3a. In order to determine the amount of volcanic SO_4^{2-} concentration, we estimate the level of non-volcanic SO_4^{2-} (i.e. background $nssSO_4^{2-}$) corresponding to the marine biogenic SO_4^{2-}. Figure 2 shows that the summer peaks of $nssSO_4^{2-}$ concentration correspond to the background $nssSO_4^{2-}$ concentration, because no explosive eruptions (VEI $\geq$ 4, mentioned later) occurred in this period, except for the 1982 eruption of El Chichón, as discussed above. We therefore determine that the background level is equal to the highest peak of $nssSO_4^{2-}$ concentration (72 ng g^{-1}) from 1982 to 1991, assuming that the background level is within this level during the past 220 years. The peaks of $nssSO_4^{2-}$ concentration exceeding this background level, shown by the thick solid line in Figure 3a, are considered to be volcanic signals.

A profile of $nssSO_4^{2-}$ concentration of the Site-J core is shown in Figure 3b. The $nssSO_4^{2-}$ concentration for the Site-J core has increased since 1860, reflecting an increase in the sulfur compound produced by combustion of solid fossil fuel and, since about 1940, liquid fossil fuel (Fujii and others, 2001). The background levels before and after about 1860 correspond mainly to the biogenic and biogenic with anthropogenic SO_4^{2-}, respectively. In contrast to the Antarctic ice cores, the background level of the Arctic cores has changed significantly, and, as the sampling interval of the Site-J core is not constant, we propose the following method for estimating the background level of the Site-J core. Firstly, we smooth the profile of $nssSO_4^{2-}$ concentration using a Stineman function which applies a geometrical weight to each datum (dashed curve in Fig. 3b). Secondly, we calculate the doubled standard deviation (2σ) of the $nssSO_4^{2-}$ concentration in the period (e.g. 1801–08) when no explosive eruptions having VEI $\geq$ 4 occurred. Then the 2σ values are plotted against age and smoothed using the Stineman function. Thirdly, we superimpose the regression curve of 2σ on that of $nssSO_4^{2-}$ concentration and we define that this curve shown by the thick solid curve in Figure 3b is the background level. Peaks of $nssSO_4^{2-}$ concentration higher than this background level are therefore considered to be volcanic signals.

Signals exceeding the background level in the H15 and Site-J ice cores are assigned to the most plausible eruptions based on the ice-core ages of the volcanic signals, the years of explosive eruptions and the volcanic explosivity index (VEI; Simkin and Siebert, 1994). The VEI proposed by Newhall and Self (1982) gives a measure of the power of eruptions, depending on such factors as volume of the erupted magma and the eruption column height. The volume of magmas logarithmically increases with increasing VEI on a scale of 0–8. All known eruptions are given a VEI value (Simkin and Siebert, 1994). The probable source eruptions are also shown in Figure 3a and b, respectively, with year of eruption and their VEIs in parentheses.

Prominent signals of the H15 core are found in two layers at depths of 44 and 46 m, which correspond to AD 1815 and 1810, respectively. Two significant peaks have been found in various cores from both Greenland and Antarctica (e.g. Dye 3 in Greenland and New Byrd in Antarctica: Langway and others, 1995), whereas the older signal is not found in the Site-J core, probably because of low time resolution. The younger signal is assigned to the eruption of Tambora, Indonesia, in 1815 with a VEI of 7. This assignment is plausible because the eruption is the most explosive during the past 220 years, producing an eruption column to a maximum height of 50 km (Rampino and Self, 1982; Self and others, 1984). A 0.7°C lowering of the mean atmospheric temperature in the Northern Hemisphere has been recorded, which probably triggered "the year of 1816 without summer" (Stommel and Stommel, 1983). The older signal has not been assigned because there are no records of probable eruptions around this date.

The most significant signal of $nssSO_4^{2-}$ concentration during the past 220 years is found at 81 m depth in the Site-J core, corresponding to AD 1784. We assign the signal to the eruption of Laki, Iceland, in 1783 with a VEI of 4. The signal of the Laki eruption is detected in various cores from the Arctic region (e.g. Summit: Zielinski, 1995, Clausen and others, 1997; Camp Century: Clausen and Hammer, 1988; Mount Logan: Mayewski and others, 1993). The eruption column height is estimated to range from 6 to 13 km (Thordarson and others, 1996). The atmospheric aerosol produced during the eruption was observed over Europe (Fiacco and others, 1994). Extreme decrease of air temperature, resulting from injection of sulfuric acid aerosols, was recorded in the Northern Hemisphere (Sigurðsson, 1982). However, this signal is not found in the H15 core, probably because the aerosol from Laki at high northern latitude did not reach the Antarctic region. Among the volcanic signals of the H15 and Site-J cores, five common signals are found, which are indicated by the dotted lines between Figure 3a and b.

BIPOLAR VOLCANIC SIGNALS

Prominent signals of the $nssSO_4^{2-}$ concentration or ECM during the past 220 years found in the ice cores from H15, Site-J and other locations in Antarctica and the Arctic are listed in Tables 1 and 2, respectively. Also given are their probable source volcanoes, years of their eruptions and VEI values. The sampling locations, except for Mount Logan, Canada, are shown in Figure 1. The age assignment of each signal differs by a few years from core to core because of dating errors and/or variability in the timing of the deposition of the volcanic aerosol onto the ice sheets. On the basis of Tables 1 and 2, common signals found from the ice cores and snow pits of the ice sheets and glaciers in the Antarctic and the Arctic regions during the past 220 years are derived from the eruptions of El Chichón in 1982, Agung, Indonesia, in 1963, Santa Maria, Guatemala, in 1902, Krakatau, Indonesia, in 1883, Cosiguina, Nicaragua, in 1835, an unknown volcano between 1831 and 1834, Tambora in 1815 and an unknown volcano in 1809.

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Table 1. Volcanic signals recorded in ice cores and snow pits in the Antarctic region during past 220 years

Year of signal (AD)	*Probable source*	*Year of eruption (AD)*	*VEI*	*Location*	*Samples*[1]
1992–1994	Pinatubo	1991	6	15.14° N, 120.35° E	DF[2], SP[2], AM
1991–92	Cerro Hudson	1991	5	46.17° S, 72.92° W	DF[2], SP[2],
1983	El Chichón	1982	5	17.33° N, 93.20° W	H15, DF[2]
1964–65	Agung	1963	4	8.34° S, 115.51° E	G15, SP, NBY, Siple, Dyer, PR, Vostok, DC, D55, D80
1905–06	Santa Maria	1912	6	14.76° N, 91.55° W	H15, DC
1886–89	Tarawera	1886	5	38.23° S, 176.51° E	H15, G15, SP, NBY, Siple, Dyer, PR
1883–85	Krakatau	1883	6	6.10° S, 105.42° E	H15, G15, SP, NBY, Siple, Dyer, PR, DC, BS, AM
1835–39	Cosiguina	1835	5	12.98° N, 87.57° W	H15, G15, SP, NBY, Siple, Dyer, PR, DC, BS
1831–34	Unknown	–	–	–	H15, SP, NBY, Siple, Dyer, BS
1815–17	Tambora	1815	7	8.25° S, 118.00° E	H15, G15, SP, NBY, Siple, Dyer, PR, Vostok, DC, D57, Siple, BS, AM
1809–11	Unknown	–	–	–	H15, G15, SP, NBY, Siple, Dyer, PR, Vostok, DC, D57, Siple, BS, AM

[1] Abbreviations of sampling sites: DF, Dome Fuji; SP, South Pole; AM, Amundsenisen; NBY, New Byrd; PR, Plateau Remote; DC, Dome C; BS, Byrd Station.

Sources: South Pole: Delmas and others (1992), Langway and others (1995), Cole-Dai and others (1997b); Dome Fuji: Kohno and others (1998); H15: Kohno and others (1999), this study; G15: Moore and others (1991); New Byrd; Langway and others (1994, 1995); Dyer: Cole-Dai and others (1997a); Siple: Legrand and Delmas (1987), Cole-Dai and others (1997a); Plateau Remote: Cole-Dai and others (2000); Dome C, D55, D57, D80: Legrand and Delmas (1987); Byrd Station: Langway and others (1995); Amundsenisen: Karlöf and others (2000).

[2] Snow-pit samples.

The 1982 eruption of El Chichón (VEI 5) is one of the most explosive in recent decades. The eruption column is estimated to have reached well into the stratosphere, 32 km high (Carey and Sigurðsson, 1986). This column height means that the eruption may have had a global impact. The amount of SO_2 injected into the stratosphere is estimated to be 7 Mt SO_2 using satellite observation (Krueger, 1983; Bluth and others, 1993). The eruption of Agung in 1963 (VEI 4) with a maximum eruption column height of 28 km (Rampino and Self, 1982; Self and King, 1996). The eruption of Santa Maria in 1902 (VEI 5) with an eruption column height of 28 km (Williams and Self, 1983). The eruption column of Krakatau (VEI 6) in 1883 rose to at least 26 km high (Mandeville and others, 1996). The 1835 eruption of Cosiguina (VEI 5) produced an eruption column 35 km high (Self and others, 1989).

These volcanic eruptions, except for the two unknown eruptions, have two common characteristics: (1) they took place at low latitudes between 20° N and 10° S, and (2) the eruption column heights exceeded 25 km (i.e. eruptions having VEI $\geq$ 5) except for the Agung eruption having VEI 4. The latter may have been a high-sulfur-release eruption among eruptions having VEI 4. The amount of SO_2 emission during this eruption is estimated to be 2.5 Mt SO_2 by a petrologic method (Self and King, 1996) which com-

Table 2. Volcanic signals recorded in ice cores and snow pits in the Arctic region

Year of signal (AD)	*Probable source*	*Year of eruption (AD)*	*VEI*	*Location*	*Samples*[1]
1982–84	El Chichón	1982	5	17.33° N, 93.20° W	SJ, 20D, Summit[2]
1970–71	Hekla	1970	3	63.98° N, 19.70° W	SJ, GISP2, 20D
1963–64	Sheveluch	1964	4	56.65° N, 161.36° E	SJ, GISP2, CR
	Agung	1963	4	8.34° S, 115.51° E	
	Surtsey	1963	3	63.43° N, 20.28° W	
1956–57	Bezymianny	1956	5	55.97° N, 160.58° E	SJ, GISP2, 20D
1947	Hekla	1947	4	63.98° N, 19.70° W	SJ, CR, 20D
1924–26	Raikoke	1924	4	48.25° N, 153.25° E	SJ, GISP2, 20D
1917–19	Katla	1918	4	63.63° N, 19.03° W	GISP2, 20D
1912–14	Katmai	1912	6	58.27° N, 155.16° E	SJ, GISP2, GRIP, CR, 20D, ML
1902–03	Santa Maria	1902	5	14.76° N, 91.55° W	SJ, GISP2, 20D
	Soufrière	1902	4	13.33° N, 61.18° W	
	Pelée	1902	4	14.82° N, 61.17° W	
1883–85	Krakatau	1883	6	6.10° S, 105.42° E	SJ, GISP2, CR, D3
1835–36	Cosiguina	1835	5	12.98° N, 87.57° W	SJ, GISP2, D3
1831–32	Unknown	–	–	–	SJ, GISP2, D3
1815–16	Tambora	1815	7	8.25° S, 118.00° E	SJ, GISP2, GRIP, CR, 20D, ML, D3, ST, CC, NC, SE, SG, SA, SB, SD
1809	Unknown	–	–	–	GISP2, GRIP, D3, ST
1783–84	Laki	1783	4	64.42° N, 17.33° W	SJ, GISP2, GRIP, CR, 20D, ML, CC, NC, SE, SG, SD, SB, SA, D2, D3

[1] Abbreviations of sampling sites: SJ, Site-J; CR, Crête; ML, Mount Logan; D3, Dye 3; ST, Site T; CC, Camp Century; NC, North Central; SE, Site E; SG, Site G; SA, Site A; SB, Site B; SD, Site D; D2, Dye 2.

Sources: 20D: Lyons and others (1990), Mayewski and others (1990, 1993); Site-J: this study; Summit: Zielinski and others (1997); GISP2: Zielinski (1995); GRIP: Johnsen and others (1992), Clausen and others (1995, 1997); Crête: Hammer and others (1980), Crowley and others (1993); Mount Logan: Mayewski and others (1993); Dye 3: Clausen and Hammer (1988), Langway and others (1995); Site T: Dai and others (1991); Camp Century, North Central, Site E, Site G, Site A, Site B, Site D, Dye 2: Clausen and Hammer (1988).

[2] Snow-pit samples in the Summit region.

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monly provides a minimum estimate (Devine and others, 1984; Palais and Sigurðsson, 1989). This emission is comparable to that of eruptions having VEI 5 based on a relation between the SO_2 emission and VEI values proposed by Bluth and others (1993). Therefore, if the eruptions having VEI 4 take place at low latitudes and are accompanied by sulfur-rich gas, the sulfuric acid aerosol may spread over and be deposited in both polar regions.

The presence of the two common characteristics mentioned above implies that the two unknown eruptions of 1831–34 and 1809 may have taken place at low latitudes, having VEI >5. These characteristics may help to narrow down the source of these unknown eruptions. One unknown signal in the period 1831–34 has been assigned to the 1831 eruption of Babuyan, Philippines, by Delmas and others (1992), Langway and others (1995) and Zielinski (1995). Volcanogenic atmospheric phenomena such as red sky, dim sun and temperature decrease, which were recorded in the early 1830s, are attributed to the aerosols of the Babuyan eruption (VEI 4) by Self and others (1989). On the other hand, Cole-Dai and others (2000) doubt the assignment because the unknown peaks found in the ice cores from Siple and Dyer in West Antarctica are separated by 2 years, suggesting that the signals do not derive from any global events. Another unknown signal in 1809–11 has been precisely resolved by Dai and others (1991). They mentioned that the signal originates from an eruption occurring in the Equatorial region in 1809, and the SO_2 emission corresponds to half of that from the Tambora eruption in 1815. No eruptions with VEI >5 were recorded during that time (Simkin and Siebert, 1994), and Palais and others (1990) find micrometer-sized andesitic glasses in the 1809 layer of the ice core from South Pole. Self and others (1989) regard the moderate eruption of Cosiguina in 1809 as a possible source of the unknown signal.

CONCLUSIONS

Common volcanic signals found in both polar ice sheets during the past 220 years are derived from the eruptions of El Chichón in 1982, Agung in 1963, Santa Maria in 1902, Krakatau in 1883, Cosiguina in 1835, an unknown volcano between 1831 and 1834, Tambora in 1815 and an unknown volcano in 1809. Volcanic eruptions potentially able to produce volcanic signals in bipolar ice sheets are characterized as follows: (1) the eruptions take place in low latitudes between 20°N and 10°S, and (2) the eruption column height reaches >25 km, mainly in the case of eruptions having VEI >5.

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