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ABSTRACT

The effect of volcanic emission of acidic aerosols on climate is well documented. The presence of acid droplets in the stratosphere can reduce transmissivity and hence decrease surface temperatures. Since the amount and chemical composition of erupted material has important effects on regional climate, knowledge of past volcanic events is of extreme importance. Detailed glaciochemical records provide the only milieu wherein the geochemistry of paleovolcanic events can be fully documented. We present a detailed sulfate and chloride record from an ice core drilled at site 20 D, 40 km SW of Dye 3 in southern Greenland. The record spans the time period 1869–1984 with chemical analyses of approximately eight samples per year. Time series decomposition and locally weighted scatter plot smoothing techniques were used to extract long term trends from the data so that individual volcanic eruptions could be documented. A number of events identified here have been unnoticed previously and a high percentage of the major chemical signatures documenting these events is associated with large decreases in temperature in the latitudinal zone 60–90 °N. Many authors have pointed out that the amount of volcanic acids such as HCl and H_2SO_4 injected into the atmosphere has a very important influence on global climate, yet this volcanic input has been difficult to quantify prior to ~1960. Our data help to alleviate this problem. These individual events can be compared to available frost tree ring data from North America, further establishing a volcanism–climatic linkage.

INTRODUCTION

The effect of volcanic emission of aerosols into the atmosphere on climate is well documented (Lamb, 1970; Pollack and others, 1976; Self and others, 1981; Rampino and Self, 1982, 1984; Sear and others, 1987, Self and Rampino, 1988). Large amounts of volcanic debris introduced into the stratosphere between 1500 and 1900 may have played a causative role in the "Little Ice Age" of this period (Lamb, 1970). Major explosive volcanic events such as Tambora (1815) and Krakatau (1883) produced a consistent but small temperature decrease on a hemispheric scale for periods up to five years (Self and others, 1981). Even smaller eruptions such as Agung (1963) produced similar temperature perturbations (Self and others, 1981). It has been suggested that volcanic emissions of sulfur and halogen aerosols may have more effect on climatic change than that of volcanic dust (Pollack and others, 1976; Rampino and Self, 1982). The presence of acid droplets in the stratosphere can reduce transmissivity and hence decrease surface temperatures. Thus, smaller sulfur- and halogen-rich eruptions may have more pronounced climatic effects than larger less halogen-rich eruptions (Rampino and Self, 1984). This may be especially true in the Northern Hemisphere (Sear and others, 1987). Since the amount and chemical composition of erupted material have an important effect on regional and global climate, knowledge of past volcanic events and of the geochemical signature of the aerosol is of extreme importance. Detailed glaciochemical records provide the best milieu wherein the glaciochemistry of paleovolcanic events can be documented.

Although other types of information are available to establish the chronology and the volume of the past volcanic events (Lamb, 1970; Simkin and others, 1981) these data sets are incomplete (Self and others, 1981). In fact, Sedlacek and others (1983) have shown through an *in-situ* stratospheric sampling program that, as late as the 1970s, many volcanic eruptions that influence the chemistry of the stratosphere went unreported. Therefore, past records of volcanic events affecting stratospheric chemistry and hence the climate are undoubtedly incomplete. In addition, there have been very few eruptions from which sufficient data exist to develop quantitative estimates of the mass of materials introduced into the atmosphere (Self and others, 1981; Devine and others, 1984).

It has long been acknowledged that volcanic emissions are an important source of several chemical species to the atmosphere. Yet it has been extremely difficult to evaluate the qualitative let alone quantitative role of volcanic emissions. Although volcanic injection of acidic anions and sulfur dioxide into the atmosphere is episodic, Sedlacek and others (1983) have shown that the volcanic contribution to the stratospheric sulfate concentration over the period 1971–81 was ~60%. In addition, stratospheric volcanic Cl^- emissions may be greatly under-estimated (Johnson, 1980). Both Neftel and others (1985) and Barrie and others (1985), utilizing ice core data from southern Greenland and Ellesmere Island, respectively, have argued that there is a strong background acid contribution to Arctic snow during the past ~80 years that cannot be accounted for by anthropogenic emissions. Previous work has demonstrated that volcanic events can be documented in Arctic ice cores (Table I).

ANALYTICAL METHODS AND PROCEDURES

In June 1984 we obtained a ~115 m core from site 20 D, ~40 km southwest of Dye 3 (65.01 °N, 44.87 °W, 2615 m a.s.l.). The top 71 m of this core has been analyzed in detail to produce an anthropogenic deposition record for SO_4^{2-} and NO_3^- in southern Greenland (Mayewski and others, 1986). Measurements of Cl^-, Na^+ and $\delta^{18}O$ concentrations were also made while selected core sections were also analyzed for NH_4^+ and F^-. Anions were analyzed using a Dionex™ Model 2010 ion chromatograph with a AS-4

TABLE I. VOLCANIC EVENTS (1870–1984) DOCUMENTED IN ARCTIC ICE CORES PRIOR TO THIS WORK

Event	Location of the sample	Technique*	Reference
Katmai, 1912	Crête	LC	Hammer, 1977
	Dye 3	SC	Neftel and others, 1985
	Hans Tavsen Ice Cap	LA, SC	Hammer, 1980
	Mt. Logan	AN	Holdsworth and Peake, 1985
	Ellesmere Island	LC	Barrie and others, 1985
Krakatau, 1983	Crête	LC	Hammer, 1977
		SC	Hammer and others, 1980
	Dye 3	AN	Herron, 1982
Bandai, 1888	Dye 3	AN	Herron, 1982
Hekla, 1947	Crête	SC	Hammer and others, 1980
Mt. Trident, 1961	Baffin Island	LC	Holdsworth, 1984
Agung, 1963	Crête	LC	Hammer, 1977
Katla, 1918	Mt. Logan	AN	Holdsworth and Peake, 1985
Mt. Wrangell, 1922	Mt. Logan	AN	Holdsworth and Peake, 1985
Raikoke, 1924	Mt. Logan	AN	Holdsworth and Peake, 1985
St. Augustine, 1934	Mt. Logan	AN	Holdsworth and Peake, 1985
Kliuchevskoi, 1937	Mt. Logan	AN	Holdsworth and Peake, 1985
St. Augustine, 1976	Mt. Logan	AN	Holdsworth and Peake, 1985

*LC = liquid conductivity, SC = solid conductivity, A = liquid acidity, AN = anionic analysis

column. Sodium was analyzed via stabilized temperature platform furnace atomic absorption spectrometry using a Perkin Elmer™ Model 2280 system. The chronology assigned to the $\delta^{18}O$ record was calibrated using known bomb layers in the total β-activity record from the upper 18 m of the core. In addition, cross-correlation of $\delta^{18}O$ from a core located 4 km to the northeast of our core site produced by the University of Copenhagen group was also used to determine depth-age relations. This glaciochemical record is one of the most detailed ever obtained with 6–8 analyses per year for a continuous 115 years. Less frequent sampling would probably not allow the observations of many of these signals.

RESULTS OF VOLCANIC INPUTS TO 20 D — 1869 TO 1984

Herron and others (1981) have shown that at Dye 3 from 1960–79 ~8% of the annual snow accumulation had melted and refrozen. However, the work of Herron (1982) and Koide and Goldberg (1985) indicates that Dye 3 snows are good preservers of volcanic records as well as of nuclear weapon testing records.

In order to reduce the potential effects of acid anion redistribution via melting/freezing we attempted to avoid ice lenses during core processing. However, the sampling of ice lenses could not always be avoided, so that the samples corresponding to years in which the percentage of ice lenses (determined from our visible observations) was greater than 10% have been eliminated from the volcanic assessment. These years are listed in Table II.

"Potential" volcanic events from site 20 D core were chosen by the criteria listed in Table III. In addition, time series decomposition and locally weighted scatter plot smoothing analysis (Cleveland, 1979; Cleveland and Terpenning, 1981) of the total excess SO_4^{2-} record with the historic trend removed provide a clear view of the signals

TABLE II. YEARS WHEN ANNUAL MELT PERCENTAGE WAS GREATER THAN 10% OF ANNUAL ACCUMULATION

1886, 1887, 1888	1943
1890, 1894, 1896	1951, 1957
1904, 1906	1969
1926	1971, 1978
1935, 1939	1982

TABLE III. CRITERIA FOR CHOOSING VOLCANIC EVENTS

Constituents	Criterion
Cl^-	>42 μgg^{-1} or double "background" of 20 μgg^{-1}
SO_4^{2-}	>52 μgg^{-1} 1869–1900 >104 μgg^{-1} 1900–60 or double "background" >222 μgg^{-1} 1960–64 or triple "background" (see Mayewski and others, 1986, for how "background" was estimated)

above the background (i.e. residuals). This approach allows a visible observation of these strong deviations from the historical trend and supports our more qualitative approach at picking volcanic peaks. *Potential* volcanic events are tabulated in Tables IV and V. The NO_3^- concentrations associated with many of these potential events (Tables IV–VI) will not be discussed here. In addition to the total Cl^- values presented in Table IV we have also presented "excess" Cl^-. This parameter is calculated like excess SO_4^{2-} using the Na^+ data and the $Na^+ : Cl^-$ ratio of seawater to calculate the amount of Cl^- present in the sample that is not associated with sea salt. The volcanic Cl^- question is complicated due to the fact that the volcanic Cl^- present could be in the form of HCl or NaCl (Woods and others, 1985). There is little doubt that certain volancoes may emit more Cl^- than So_4^{2-} during their explosive stage (Devine and others, 1984).

Volcanic loading is dependent on, in addition to other variables, source strength and the duration of the eruption. Hammer (1977) has suggested that it is possible to have a one-year lag between a volcanic eruption south of 50°N and the subsequent deposition of its aerosol products in Greenland. Two- and three-year lags have been shown for near-equatorial volcanic events (Herron, 1982). In addition to these considerations, volcanic events south of 20°S are unlikely to be observed in Greenland ice (Hammer, 1977; Hammer and others, 1980).

Although volcanic emissions of Cl^- and F^- can be quantitatively important to the global atmosphere (Cadle, 1980; Symonds and others, 1988), only through stratospheric emission and/or through tropospheric transport from a North American volcano is it likely that substantial amounts of

TABLE IV. VOLCANIC Cl^- SIGNALS AT 20 D

Year	Time of year*	Cl^- ($\mu g\ kg^{-1}$)	Excess Cl^-	Other excesses	Event
1984	winter/spring	260	0	SO_4^{2-}	El Chichon
1981	spring/summer	55	0	–	Mt. St. Helens, Garelo, Ulawu
1979	winter/spring	46	8	–	major undocumented stratospheric injection
1978	winter/spring	51	6	–	Benzymianny
1963	spring	93	18	–	Surtsey
1959	winter/spring	190	5	–	Chikurachki-Tatan
1954	winter/spring	57	0	NO_3^-	Mt. Spur
1950	fall	143	0	–	?
1947	winter→summer	53	11	–	Hekla
		180	29	F^-	Hekla
		60	11	–	Hekla
		48	10	–	Hekla
1945	winter/spring	47	11	–	Cleveland?
1942	fall/winter	110	1	–	Cotopaxi, Paulo
1941	fall/winter	190	0	–	?
1941	spring	51	23	SO_4^{2-}	?Grimsvötn
1936–37	winter	63	6	–	Asama, Magesh, Kliuch
		56	8	–	
1936	summer	63	6	–	Asama
1933	fall/winter	80	13	–	Fuego, Quiza
1930–31	fall/winter	52	14	–	Sarych
		64	12	SO_4^{2-}	Palmch? Sarych
1930	spring	400	383	–	Mt. Peleé or Ko?
1929	spring	52	13	–	Palmeh
1928	spring	100	28	–	Multiple activity Kuryils, Japan
1925	fall/winter	49	4	–	Galeras? or
		48	8	–	Raikoke?
		47	8	–	
1922	summer/fall	64	5	–	Mt. Wrangell
		50	4	–	
1921	winter/spring	46	6	SO_4^{2-}	?Manam
1920	winter/spring	71	9	–	Kumagatake, Asama
1917	spring/summer	283	2		?
1913	winter/spring	54	5		Katmai, 1913 eruption near Hekla
1911	winter/spring	115	7		Japan, Aleutians, Kamchatka
1906	winter/spring	101	0		Japan
1903	fall	120	0	SO_4^{2-}	Santa Maria, Thordarhyn
1902	fall/winter	53	0		Mt. Peleé, Soufrière
1901	winter/spring	53	0		Dona Juana? Adatano
1898	winter/spring	100	0		Mayon?
1896	fall/winter	47	0		Bandai? Bogoslof
1890	winter/spring	84	0		Shiretoko Iwo Zan
1885	winter/spring	65	0		Krakatau?
1880	winter/spring	81	7		?
		50	2		?
1879	winter/spring	102	0		Cotopaxi?, 1878 eruption near Hekla
		238	29	SO_4^{2-}	?
1876	fall	65	5		?
1874	fall	130	35		Tarumai?
1873	winter/spring	89	15	SO_4^{2-}	Grimsvötn
1871	winter/spring	66	13	SO_4^{2-}	Cotopaxi?

Volcanic events taken from Hoyt, 1979; Simkin and others, 1982; Sedlacek and others, 1983.

* From $\delta^{18}O$ profiles.

volcanic Cl^- and F^- would reach Greenland. Recent work using *in-situ* techniques have established volcanic stratospheric inputs of both HCl and NaCl (Lazrus and others, 1979; Woods and other, 1985). This is an important consideration in establishing volcanic Cl^- records in ice cores. In only two events is the excess $Cl^- \approx 50\%$ of the total Cl^- measured. These events are the 1930 and the 1941 Cl^- spikes. We have interpreted the 1941 spike to be the eruption of Grimsvötn, Iceland, while the 1930 event is unknown. The 1930 spike represents a massive injection of excess Cl^- onto the southern Greenland ice sheet. It is curious that no extremely large eruption is documented in late 1929 or early 1930 to explain this signal. In addition, the Cl^- peak is preceded by a very large NO_3^- signal in late 1929 (Table VI). There is no associated excess SO_4^{2-} event. It is possible that these peaks are due to the Komaga-Take eruption in Japan in 1929. However, 1947 Hekla and 1963 Surtsey have only ~20% of their total Cl^- recorded at 20 D as excess Cl^-. This would suggest that the 1930 event could have been a nearby Icelandic eruption, for it appears that the majority of volcanic HCl reaching southern Greenland is usually neutralized during transit. It is probable that much of the stratospheric gaseous Cl^- is also neutralized in transit to southern Greenland.

TABLE V. VOLCANIC SO_4^{2-} SIGNALS AT 20 D

Year	Time of year*	Excess SO_4^{2-} ($\mu g\ kg^{-1}$)	Other excesses	Event**
1984	winter/spring	229	Cl^-	El Chichon
1968	summer/fall	308		Redoubt, Mt. Trident
1867	fall/winter	236		Stromboli/Suwanosejima
1957	winter/spring	115		Bezymianny
1955	winter	128		?
1953	spring/summer	107		Tao-Rusyr, Mt. Trident
1950	spring/summer	123	SO_4^{2-} first, then	?
		135	NO_3^-, then Cl^-	
1947	summer/winter	110	Cl^- first, then	Hekla
		130	SO_4^{2-}	
1945	spring	110	Cl^- first, then NO_3^-, then SO_4^{2-}	Cleveland?
1944	fall	140		Vesuvius?
1942	spring	133	SO_4^{2-} first, then Cl^-	Colima, Cotopaxi? Pavlof
1941	spring	232	Cl^-	Grimsvötn?
1932	spring/summer	238		Kliuchevskoi
1931	winter	118	SO_4^{2-}	Paluweh Sarychev, Mt. Peleé?
1927	spring	126		Tokachi?
1925	winter/spring	128	SO_4^{2-} first, then Cl^-	Raikoke
1919	winter/spring	116		Katla
		159		
1914	winter/spring	116	SO_4^{2-} first, then	Katmai
		106	Cl^-, then SO_4^{2-}	
1913	winter/spring	157		Katmai
		140		
1912	fall	160		Katmai
		119		
1903	spring	119	Cl^- first, then	Mt. Peleé, Soufrière,
	fall	119	SO_4^{2-}	Thordarhyrna, Santa Maria?
1900	summer	160	SO_4^{2-} first, then Cl^-	Dona Juana
1899	spring	56	Cl^- first, then SO_4^{2-}	Mayon?
1894	summer/fall	369		melt? or
		140		major unknown
1893	summer	75		Seguam?
1888	summer	58		melt? or Bandai
1886	summer	119		melt? or Tunguarahua
1879	spring/summer	157	Cl^- first, then	Cotopaxi?
		60	SO_4^{2-}	
		54		
		52		
1878	winter/summer	79		?
		58		
		67		
1878	fall	112	Cl^-	?
	spring	60		
1875	summer	87		Askja
1874	fall	61	Cl^- first, then SO_4^{2-}	Tarumai?
1873	winter	74	Cl^-	Grimsvötn
		61		
1871	winter	53	Cl^-	Cotopaxi?

** volcanic events taken from Hoyt, 1979; Simkin and others, 1982; Sedlacek and others, 1983.

* from $\delta^{18}O$ profiles

The major Cl^- events observed at site 20 D include the Hekla, Grimsvötn and 1929–30 eruptions mentioned above, as well as Cotopaxi (1878), Agrigian (1917), Chikurachki (1959) and the most recent El Chichon eruptions (Mayewski and others, 1987). There is also an undocumented 1950 event.

Fluoride measurements have also been undertaken on many of these samples. Only in the two core segments was F^- detected as greater than 1 $\mu g\ kg^{-1}$. Our results indicate that in only a very few limited cases does HF contribute to the volcanic input of acid to southern Greenland during the period 1869–1984. Our maximum F^- concentrations are much lower than were previously reported by Herron (1982). Only after the eruptions of Hekla (1947) and Katmai (1912), the two most obviously reconizable volcanic events in the 115 year record (Tables IV–VI), do we measure any F^-. The major SO_4^{2-} signals include the 1894, 1941, 1967, 1968, and 1984 events. (The 1984 has been previously interpreted as El Chichon by Mayewski and others, 1987.)

DISCUSSION

It is clear from this work and that of Holdsworth and Peake (1985), Delmas and others (1985a) and Legrand and Delmas (1987) that only through detailed anionic analyses along with precise and accurate dating can volcanic records such as these be produced from ice cores. The events with both Cl^- and SO_4^{2-} peaks and in some cases Cl^-, SO_4^{2-} and NO_3^- peaks are indeed volcanic signatures. These eruptions are listed in Table VI.

TABLE VI. 20 D VOLCANIC EVENTS

Both Cl^- and SO_4^{2-} events		Cl^-, SO_4^{2-} and NO_3^- events	
1871	?	1879	Cotopaxi?
1874	Tarumai?	1917	Agrigan?
1876	Askja?	1945	Cleveland?
1898–99	Mayon?	1950	?
1900	Dona Juana		
1903	Mt. Peleé, Soufrière Thordarhyrna		
1912–14	Katmai, Hekla		
1925	Raikoke		
1931	Paluweh?		
1941	Grimsvötn?, or low-latitude unknown event (see text)		
1942	Colima? Cotopaxi?		
1947	Hekla		
1954	Mt. Spurr		
1984	El Chichon		

Holdsworth and Peake (1985) have also observed the eruptions of Katla (1918–19), Raikoke (1924–25) and Hekla (1947) via detailed anion analysis from an ice core from Mt. Logan, Canada. These workers have also detected eruptions that are not clearly evident from our data. They include Mt. Wrangell (1922), St. Augustine (1934–35) and Kliuchevskoi (1937–38).

Are there any other data to support our contention that these chemical signals are volcanic events? Although not explicit in their work, data sets of both Neftel and others (1985) and Barrie and others (1985) have corroborated many of our volcanic signals. In the Neftel and others (1985) data from Dye 3, years when the averaged SO_4^{2-} data lie well above their spline-smoothed curve are shown in Table VI. For the Barrie and others (1985) data from a core from Ellesmere Island, Canada, we have taken their conductivity data for years that show two peaks per year instead of the usual single peak per year (i.e. one of these peaks is presumed to be volcanic), as well as years with extremely high single peaks. (This was done by comparing peaks to adjoining year data.) These years are also listed in Table VII. It cannot be circumstantial that there is good agreement among the different data sets. Although all the data are not always in exact agreement (i.e. ±1 year), this comparison of data sets supports our contentions of volcanic events being recorded in glacier ice and snow throughout the Arctic region.

TABLE VII. COMPARISON OF VOLCANIC EVENTS BETWEEN OUR WORK AND THAT OF NEFTEL AND OTHERS (1985) AND BARRIE AND OTHERS (1985). (SEE TEXT FOR EXPLANATION)

Our work	Neftel and others	Barrie and others
1903 – Mt. Peleé, Soufrière	1903	ND
1906 – melt event?	1906	ND
1912–14 – Katmai	1912–13	1912
1917 – ?	1917	1916
1925 – Raikoke	1924	1924–25
1931 – Paluweh?	1930	1931, 1933?
1941 – ?	1941	1940
1942 – ?	1943	1943

ND = No data

Other types of information such as atmospheric transmission measurements at Table Mountain, California (the Smithsonian Astrophysical Observatory) indicate that such Southern Hemispheric eruptions as Paluweh (1928) and Quizapu (1932) were easily observed in the mid-latitudes of the Northern Hemisphere (Hoyt, 1979). These eruptions can also be documented in the 20 D core (Tables IV–V).

Notably there are eruptions that we cannot document at site 20 D which one might expect to have been recorded. For example, we do not detect Agung (1963). Interestingly enough neither Delmas and others (1985b) nor Koerner and Fisher (1982), respectively, observed it at Mt. Logan and the Agassiz Ice Cap, Canada. Eighty per cent of the volcanic debris injected by the Agung eruption remained in the Southern Hemisphere (Delmas and others, 1985a; Self and Rampino, 1988). Self and Rampino (1988) have recently argued that the estimate by Hammer and others (1980) of Agung fallout in southern Greenland is too high. Our data support this contention. Our data interpretation also agrees with their idea that some of the acid measured by Hammer and others (1980) in 1963 could be related to the Surtsey eruption (see Table IV, especially excess Cl^-). Although we observe what could possibly be a small Cl^- excess due to the Krakatau eruption (1883) (Table V), its significance in the 20 D record is small. Delmas and others (1985a) have argued that both Agung and Krakatau should probably not be detected in Greenland ice. Our data also show little to no effect in southern Greenland of the Mt. St. Helens (1980) eruption. We observe a small Cl^- signal (Table IV) but no SO_4^{2-} signal. This is not surprising because this eruption was very poor in H_2SO_4 but relatively rich in halogens. As Self and Rampino (1988) state, "in terms of climate Mt. St. Helens was a non-event". Our data support this premise.

CLIMATIC IMPLICATIONS

Probably the most significant aspect of this work is the climatic implications. Volcanic aerosols can be major contributors to regional and global temperature changes both in the long- and short-term (Lamb, 1970; Pollack and others, 1976; Bradley and England, 1978; Self and others, 1981; Rampino and Self, 1982, 1984; Stothers, 1984). Eruptions such as Tambora (1815), Krakatau (1883), Santa Maria (1902), Katmai (1912), and Quizapu (1932) have produced temperature decreases on the order of 0.2–0.5 °C on a hemispheric scale for time periods up to five years (Self and others, 1981). The temperature perturbations due to volcanic aerosol emission may reach as high as 1.5 °C (in earth surface temperatures) in the high-latitude zones (Self and others, 1981).

The average yearly ΔT° from 1875 to 1977 in the latitudinal zone 60°–90 °N tabulated by Self and others (1981) is shown in Figure I. Their arrows indicate the eruptions of Krakatau, Santa Maria + Peleé + Soufrière, Katmai, Hekla, Bezymianny and Agung were followed by rapid decreases in ΔT°. Using the data presented within, not only do the Santa Maria + Peleé + Soufrière, Katmai and Hekla volcanic signals documented in southern Greenland correspond to ΔT drops, but at least eight other volcanically associated ΔT decreases can also be observed (Fig. 1). The ΔT° decrease Self and others (1981) have associated with Bezymianny might actually be associated with the Mt. Spurr eruption of 1954. Many of the decreases in ΔT between 1884–85 to 1902 may also be volcanically induced but our data from this period are not as convincing (i.e. only Cl^- or SO_4^{2-} spikes, not both or the dates of ΔT° drops and Greenland ice core signals are offset by a year or more). The decrease of ΔT° after 1960 may also be volcanically related but the SO_4^{2-} anthropogenic record probably makes it. This dramatic coincidence of ΔT° drop above 60 °N latitude and volcanic acid input observed in southern Greenland snow and ice is probably not coincidental. It is our supposition that the volcanic eruptions documented in the ice core indicate a possible cause and effect relationship resulting in small but significant temperature decreases.

The argument is further substantiated by measured decreases in $\delta^{18}O$ of snow after the deposition of volcanic

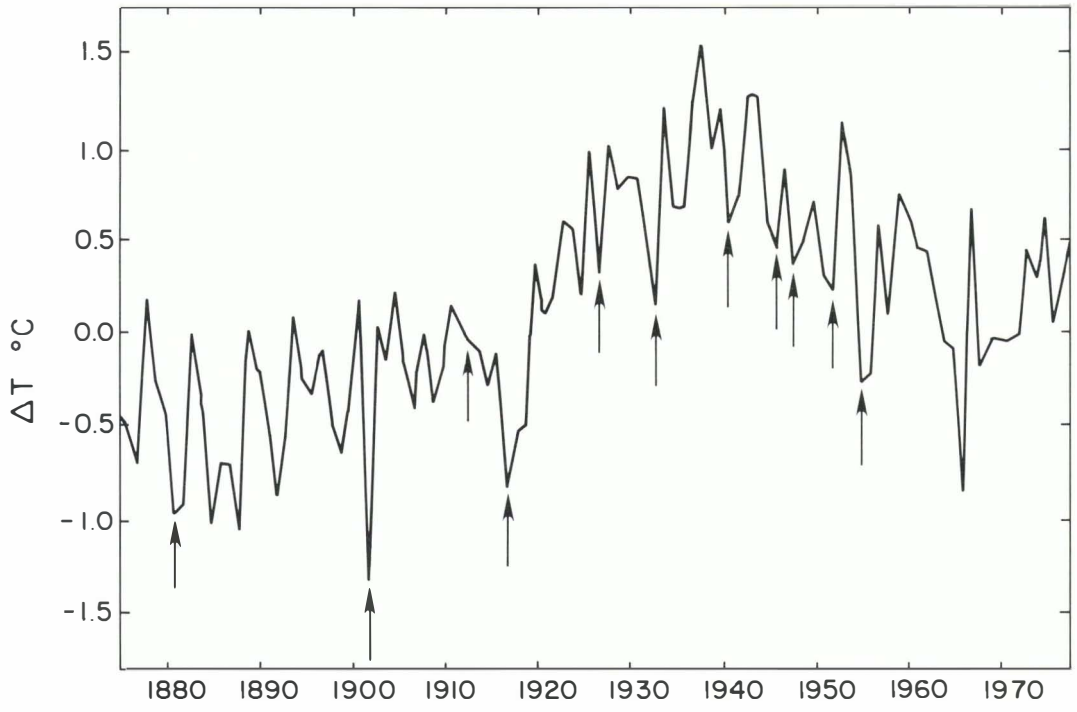

Fig. 1. Average yearly temperature change between 60–90°N from 1875 to 1977; from Self and others (1981). Arrows present volcanic events observed in 20 D ice.

debris both at our 20 D site (Mayewski and others, 1987) and at Mt. Logan (Holdsworth and others, 1986). These more negative $\delta^{18}O$ signals in the precipitation imply an immediate local to regional temperature decrease upon the "arrival" of the volcanic aerosol, such as that from the recent El Chichon eruption (Mayewski and others, 1987). This "event" could have produced a 0.5°C cooling event during the summer of 1983 in southern Greenland based on oxygen isotope–air temperature calibration (Davidson and others, 1987).

The previously undocumented 1941 event bears some detailed discussion. Although many frost tree ring events in the western USA have been associated with large volcanic events, one of the four major frost tree ring events in this century was in 1941. It has heretofore not been correlated to a *known* volcanic event (LaMarche and Hirschboeck, 1984). It is curious that one of the largest volcanic "events" observed by us at site 20 D was during this time period. Could this large event have gone unreported due to pre-occupation with World War II, as previously suggested by Simkin and others (1981)?

Handler (1986) has recently shown a strong association between stratospheric aerosols and Indian monsoon precipitation. He found that low-latitude aerosols precede below-average precipitation and high-latitude aerosols precede above-average precipitation. He showed that in 1941 there was below-average monsoon precipitation that has not been correlated to a specific volcanic event. Using his arguments and models, the volcanic event recorded by us at 20 D and in the frost tree ring records of the western USA by LaMarche and Hirschboeck (1984) was probably a low latitude (0–25°N) volcanic eruption. In addition to the 1941 event, our previously unrecorded volcanic events can be utilized to support Handler's (1986) contentions regarding monsoonal variations. Our 1917 and 1942 "events" (Table VI) are probably high-latitude eruptions because they coincide with above-average monsoonal precipitation (Handler, 1986). This is not to say that all monsoonal variations can be interpreted in light of our volcanic data, but that in many cases they can be.

We feel strongly that the technique reported here and in Legrand and Delmas (1987) present the most significant means for detailed quantification of individual volcanic events in the past. In turn, this information can then be used to compare to other historic climatic data in order to assess accurately the role of volcanism on climate.

CONCLUSION

The results of our glaciochemical measurements conducted on firn and ice samples from southern Greenland indicate that many volcanic eruptions of Northern Hemispheric as well as of global concern are clearly recorded by elevated concentrations of Cl^-, SO_4^{2-} and in some cases also NO_3^-. Although the majority of these acidic anions has been transported via the stratosphere, Delmas and others (1985a) have argued that due to the closeness of many volcanic centers to Greenland the volcanically produced anions could also be transported in the troposphere. Our data show a strong coincidence of volcanic anion deposition and temperature drops within the latitude band 60–90°N for individual volcanic events.

More work of this type in various locations in the Arctic would provide valuable information concerning the type, strength and transport mode of volcanic eruptions. These records could then be better correlated to known climatic records and allow the development of better climatic models. This is particularly true in attempting to assess the role of relatively small but sulfur-rich eruptions (Self and Rampino, 1988).

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