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Resolution of environmental variation by detailed analysis of YM85 shallow ice core in Antarctica

Hirotake TAKAHASHI[1], Taeko YOKOYAMA[1], Makoto IGARASHI[2], Hideaki MOTOYAMA[3] and Keisuke SUZUKI[1,4]
1 Department of Environmental Sciences, Shinshu University, Matsumoto 390–8621, Japan
2 RIKEN, Wako, Saitama 351–0198, Japan
3 National Institute of Polar Research, Tachikawa 190–8518, Japan
4 Institute of Mountain Science, Shinshu University, Matsumoto 390–8621, Japan

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Abstract

An ice core with a length of 77.71 m water equivalent (w. e.) obtained in January 2002 from YM85 in East Dronning Maud Land, Antarctica, has been subjected to detailed analysis. The ice core is dated at 767 years (1233–1999 A.D.) on the basis of the tritium concentration and volcanic signal time markers and by annual layer counting. Large volcanic eruptions such as those of Tambora in 1815 and Kuwae in 1453 are likely to have caused the decrease in the temperature at YM85. During 1453–1816 A.D., the surface mass balance at YM85 was significantly lower than that observed prior to and subsequent to this period. This is thought to be because of LIA (little ice age) and the hiatus in snow deposition. Observation of the annual surface mass balance from 1816 A.D. to 1998 A.D. shows that it decreases sharply from the 19th century to the 20th century. The correlation coefficient between the five-year average values of δ^{18}O and those of the annual surface mass balance is positive (R=0.62) for the period 1816–1998 A.D. The decrease in the Na^+ concentration with an increase in the temperature is inferred to be a consequence of the diluting with a large amount of snowfall.

1. Introduction

Japanese Antarctic Research Expeditions (JAREs) have collected shallow ice cores from East Dronning Maud Land, Antarctica, in order to reconstruct the paleoenvironment. The reconstruction of the paleoenvironment is a part of the International Trans-Antarctic Scientific Expedition (ITASE) program, which is an international effort to obtain paleoclimate records and clarify the environmental variations during the last few hundred years in Antarctica (Kameda *et al.,* 2002). Under these plans, a shallow core is drilled at YM85, Antarctica.

Ice sheets are formed by the deposition of snow over a long period, and the reconstruction of past variations in the atmospheric environment on the basis of the analysis of ice cores is thought to be possible. The ice cores are expected to contain information on the temperature variation by stable isotopic ratio, volcanic eruptions, surface mass balance etc., in the past. It is also supposed to contain records of seasonal variations in the concentrations of chemical components that can help in the determination of the origins of these chemical components. It is important to clarify how these environmental factors interact with each other in the process recorded in ice cores in resolving the environmental variations.

The objective of this study is to investigate the relationship between temperature variations and other environmental factors at YM85 by performing a detailed analysis of the shallow ice core obtained from the location. We also report the result of the dating of the ice core.

2. Methods

2.1 Drilling site

In January 2002, a 105.51 m ice core was collected from the YM85 (71°35′14″S, 40°38′5″E; 2246 m a.s.l.) in East Dronning Maud Land, Antarctica (Fig. 1) by JARE-42. This site is located in the Shirase Glacier, at the midpoint between Mizuho Station and the Yamato Mountains. Sastrugi are often observed around the YM85; hence, this area is thought to be characterized by strong katabatic winds. The average flow speed of the surface ice sheet is 25.5 m year^{-1} northward and 5.8 m year^{-1} westward (Motoyama *et al.,* 2002).

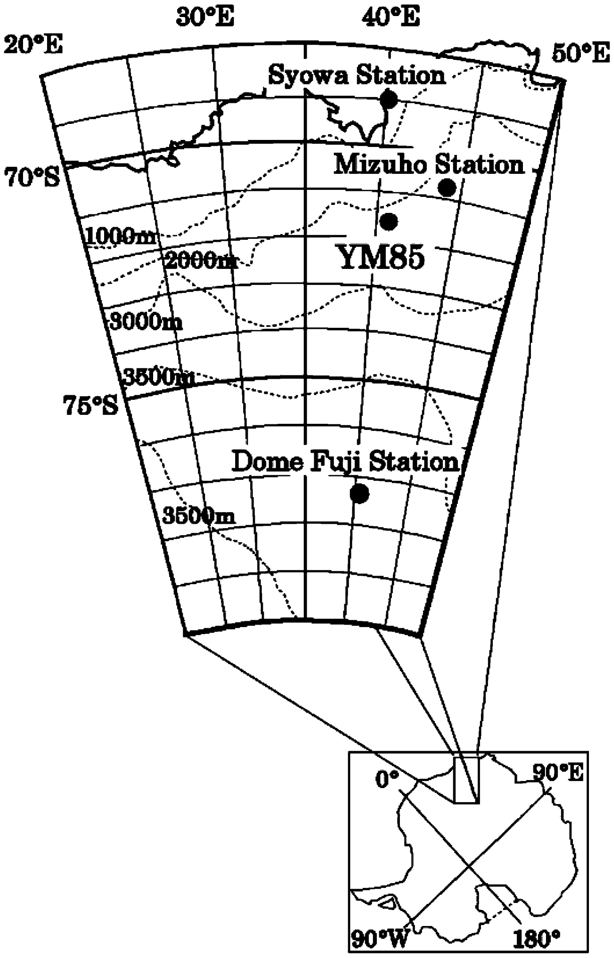

Fig. 1. Location of the YM85 shallow ice core drilling site, East Dronning Maud Land, Antarctica.

2.2 *Analysis*

The YM85 shallow ice core was transported to Japan in a cold storage. The core is then cut into pieces with lengths of 2 to 6 cm in a low-temperature laboratory so that every piece provides at least 20 ml of liquid upon being melted. Subsequently, the surface of the ice core was scraped with a ceramic knife washed in ultrapure water to remove contaminants. The decontaminated sample was melted in a Teflon bottle washed with ultrapure water in a clean room for melting, and the resulting liquid sample was poured into a plastic bottle washed with ultrapure water, and the plastic bottle was refrigerated until the beginning of chemical analysis. We used the melted sample to perform chemical and isotope analyses. After electrical conductivity (EC) and pH of the samples were measured, the concentrations of the major chemical components (Na^+, NH_4^+, K^+, Mg^{2+}, Ca^{2+}, F^-, MSA^- , Cl^-, SO_4^{2-}, NO_3^-)were measured by ion chromatography (anions: DIONEX DX-320, cations: DIONEX ICS-1500). We determined the isotopic composition of oxgen-18 ($\delta^{18}O$) using a mass spectrometer (THERMO FISHER Delta Plus) on CO_2 equilibrated with samples. Detailed analysis methods are described in Watanabe *et al.* (1997) and Igarashi *et al.* (1998). The liquid scintillation method is used for the determination of tritium content in the liquid phase. Detailed analysis method of tritium is described in Kamiyama *et al.* (1997).

2.3 *Conversion of the depth of the ice core to water equivalent depth*

The ice core is composed of firn and ice, and the density varies considerably along the core; the densities at the top and bottom of the core are 0.43 g cm^{-3} and 0.91 g cm^{-3}, respectively (Motoyama *et al.,* 2002). The depth of ice core must be converted into the water equivalent depth (w. e.) scale using the density data along the core. The water equivalent length of the sample is obtained by multiplying the density of the ice core by the length of the sample, and the total water equivalent depth of the ice core is obtained by the summation of the water equivalent lengths of all the samples. For layers whose density data is unavailable, the average density of the top and bottom layers is used. The water equivalent depth at the bottom of the ice core is 77.71 m w. e.

3. Results and Discussion

3.1 *Chemical profile of the YM85 ice core*

Figure 2 shows the fluctuations in the value of $\delta^{18}O$, the concentrations of Na^+, MSA^-, non-sea-salt sulfate ($nssSO_4^{2-}$) and NO_3^-. The value of $\delta^{18}O$ in the surface snow is associated with the 10 m snow temperature (Satow and Watanabe, 1992), and hence, we can use the fluctuation in $\delta^{18}O$ in the core as an indicator of the temperature. In the approximate depth ranges of 18–22 m w. e. and 36.5–38 m w. e., the fluctuations in $\delta^{18}O$ and $nssSO_4^{2-}$ concentration appear to be significantly high, while MSA^- concentration appears to be significantly low (Fig. 2).

The NO_3^- concentration also shows a significant decrease in the depth ranges of 18–22 m w. e. and 36.5–38 m w.e. (Fig. 2). The NO_3^- concentration has been observed to decrease with an increase in temperature in ice cores obtained from Greenland and Antarctica, and the post-depositional loss of NO_3^- from the top snow layers is caused by the release of HNO_3 and the photolysis of NO_3^- (Rothlisberger *et al.,* 2002). The large decrease in the NO_3^- concentration in the approximate depth range of 20–37 m w. e. can be attributed to the post -depositional loss since a significant increase in $\delta^{18}O$ is observed at these depths.

Comparing the fluctuation in $\delta^{18}O$ with that in Na^+ concentration, it is found that $\delta^{18}O$ is inversely proportional to the Na^+ concentration (Fig. 2). It is suggested that the source of Na^+ in the ice core is sea salt (Fig. 3), and Na^+ concentration is expected to be high during cold climatic conditions (Suzuki *et al.,* 2005); however, observations show that the reverse is

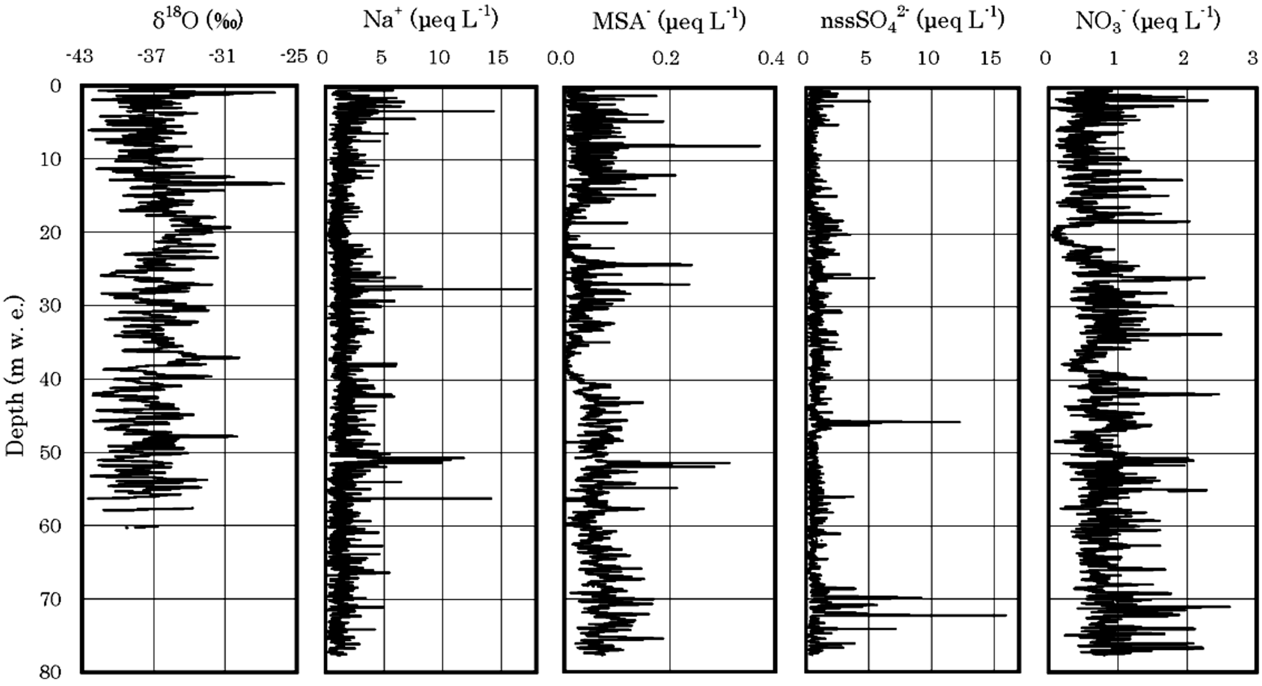

Fig. 2. Vertical profiles of chemical components of the YM85 ice core.

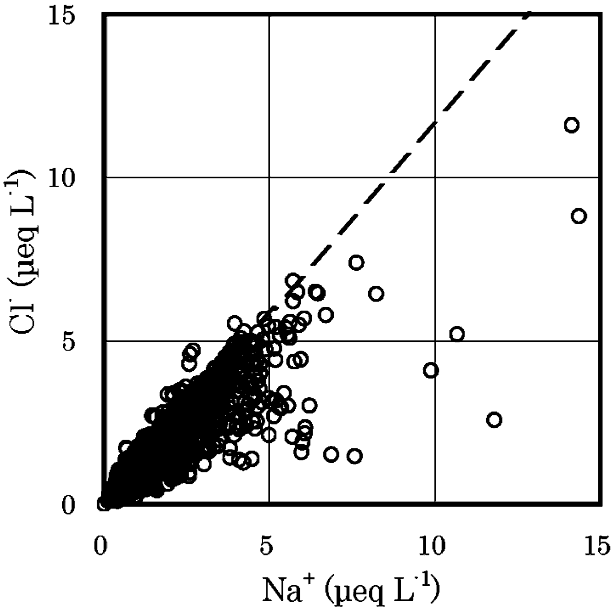

Fig. 3. The relationship between concentrations of Na^+ and Cl^- in YM85 ice core. The broken line in this figure indicates the ratio of concentrations of Na^+ and Cl^- in seawater.

true. In section 3.4.2, using the surface mass balance information obtained from the ice core, we shall explain the relationship among the fluctuations in $\delta^{18}O$, the Na^+ concentration, the MSA^- concentration, and the NO_3^- concentration.

3.2 Dating of the YM85 ice core

As a result of snow pit work at the drilling site, the annual surface mass balance for the last two to three years is estimated to be about 200 mm w. e. $year^{-1}$ (Toyama *et al.,* 2005), and hence, it could be thought that annual layer has several samples. However, the annual surface mass balance is not always constant for long periods of time throughout the ice core. We attempted to identify the annual layers as follows: first, we employed several highly reliable time markers. We then counted the number of annual layers using several ionic concentrations believed to show seasonal variations.

3.2.1 Time markers

The first time marker is the tritium concentration (Fig. 4). During the early 1950s and 1960s, the testing of thermonuclear atmospheric bombs resulted in the emission of a large amount of artificial radioactive nuclides into the earth's atmosphere, and hence, annual layers with remarkably high tritium concentrations are observed in the Antarctic ice cores (Stenni *et al.,* 1999; Nishio *et al.,* 2002). The graph of tritium concentration vs. the depth for the ice core is shown in the basic analytical report of the YM85 shallow ice core. The graph shows large peaks at depths of 4.84, 4.19, and 5.02 m w. e., corresponding to the years 1966, 1969, and 1965, respectively (Igarashi *et al.,* 2003). We set these three tritium peaks as the first time markers.

The second time marker is the signals generated by large volcanic eruptions. Large volcanic eruptions release large amounts of sulfur oxides into the atmosphere. These oxides are eventually precipitated on the polar ice sheet surface. Hence, the presence of large amounts of $nssSO_4^{2-}$ in the Antarctic shallow ice cores can be attributed to the volcanic eruptions (Kohno *et al.,* 1999; Stenni *et al.,* 1999; Kohno, 2000; Nishio *et al.,* 2002; Suzuki *et al.,* 2005; Zhou *et al.,* 2006). A plot of the nssSO_4^{2-} concentration against the depth of the YM85 ice core shows large peaks at depths of 26.1 m w. e., 45.8 m w. e., and 72.3 m w. e. (Fig. 2), and the fluctuation pattern of these peaks is similar to that of the nssSO_4^{2-} concentration in the DT263 ice core (Zhou *et al.,* 2006). According to Zhou *et al.* (2006), these peaks correspond to the volcanic eruptions of Tambora in 1815, Kuwae in 1453 and Unknown in 1259, respectively. By taking into account the time taken for the transportation of the oxides from the site of the volcanic eruptions to the Antarctic ice sheets, we estimated the deposition of nssSO_4^{2-} on the layer at depths of 26.1 m w. e., 45.8 m w. e., and 72.3 m w. e. to have occurred in 1816, 1454, and 1260 A. D., respectively. We set these three years that correspond to the years of generation of volcanic signals as the second time markers.

Table 1 shows the mean annual surface mass balance calculated for the period between the time markers mentioned above. The mean annual surface mass balance from the surface snow layer to the layer formed in 1816 A.D. and from the layer formed in 1454 A.D. to the layer formed in 1260 A.D. is in the range of 127 mm-217 mm w. e. year^{-1} (Table 1), and we could observe several annual layers corresponding to the above range. On the other hand, the mean annual surface mass balance from the layer formed in 1454 A.D. to the layer formed in 1816 A.D. is about 54 mm w. e. year^{-1} (Table 1), we could not observe many annual layers corresponding to this range. We attempted to identify the annual layers by assuming that the annual surface mass balance was constant from 1454 A.D. to 1816 A.D. In section 3.4.1, we shall explain the reason for the decrease in the surface mass balance during this period.

3.2.2 *Identification of annual layer*

We use the seasonal variations in the concentrations of MSA^-, NO_3^-, and Na^+ in order to identify the annual layers of the ice core. As described above, the source of MSA^- is marine phytoplankton; this implies that the MSA^- should be high in summer and low in winter (Suzuki *et al,* 2005). It is thought that NO_3^- contained in the Antarctic ice core mainly derived from the stratosphere (Stenni *et al.,* 1999); the nssSO_4^{2-} and NO_3^- concentrations indicate high peaks in summer (Li *et al.,* 1999; Nishio *et al.,* 2002). As described above, sea salt is the source of Na^+ (Fig. 3); this implies that the Na^+ concentration should be high in winter and low in summer (Li *et al.,* 1999; Zhou *et al.,* 2006). It is suggested that the maximum MSA^- concentration nearly corresponds to that of the maximum NO_3^- concentration in the layer formed during summer, and this layer is considered as the boundary

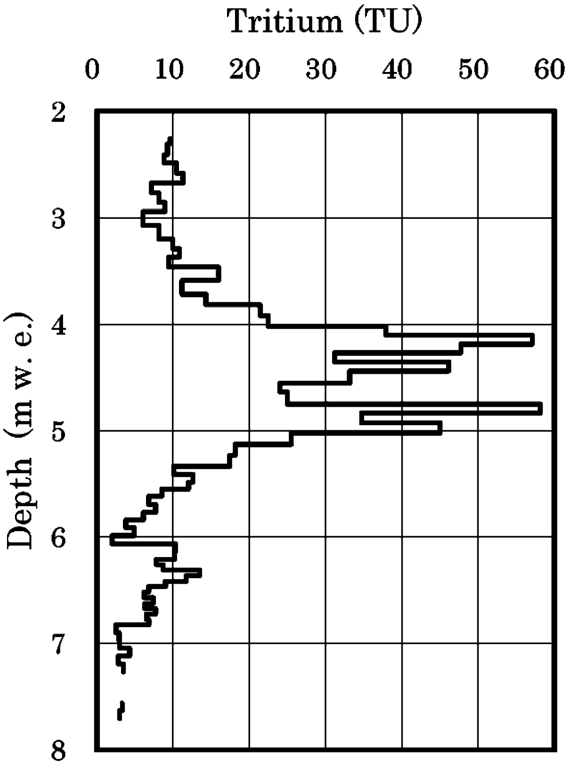

Fig. 4. The fluctuation of tritium concentration in the YM85 shallow ice core.

Table 1. Mean annual surface mass balance calculated between tritium and large volcanic time markers in YM85 ice core.

Time markers	Water equivalent depth (m w.e.)	Time in core (A.D.)	Mean surface mass balance (mm w.e.yr^{-1})
Surface snow layer	0	2002	
			127
Tritium	4.19	1969	
			217
Tritium	4.84	1966	
			187
Tritium	5.02	1965	
			141
Tambora	26.10	1816	
			54
Kuwae	45.78	1454	
			137
Unknown	72.29	1260	

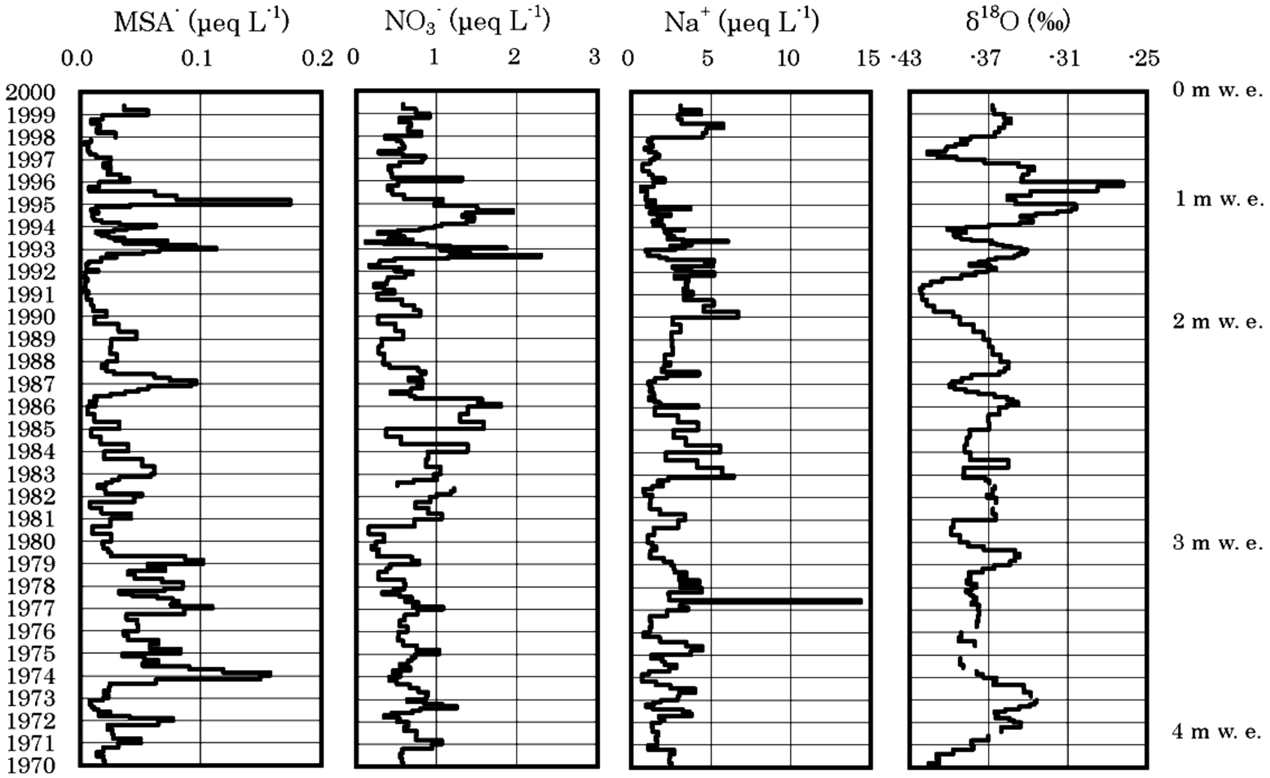

Fig. 5. Profiles of the concentrations of MSA^-, NO_3^-, Na^+, and $\delta^{18}O$ value from the surface layer to 1970 A.D.

of the annual layer (Fig. 5). When the MSA^- concentration is significantly low at a depth of around 20 m w. e. (Fig. 2), the layer with the maximum NO_3^- concentration corresponds to that of the minimum Na^+ concentration is considered as the boundary of the annual layer. We corrected the time obtained from the annual layer counting by using the tritium concentration and volcanic signal time markers described above.

Using the time markers and by annual layer counting, the annual layer at a depth of 77.71 m w. e. in the YM85 ice core is dated at 1233 A.D.

3.3 Signals generated by volcanic eruptions

3.3.1 Estimation of volcanic signals

Following the approach that is widely employed, the $nssSO_4^{2-}$ concentration values exceeding the average ($\bar{x}$) plus 2σ (σ: standard deviation) are considered as volcanic signals (Nishio *et al.,* 2002; Suzuki *et al.,* 2005; Zhou *et al.,* 2006). However, in the present study, it is supposed that only a small number of volcanic signals are observed because the background fluctuation in the $nssSO_4^{2-}$ concentration is added to the effect of large volcanic eruptions. The average ($\bar{x}'$; 0.65 μeq L^{-1}) and standard deviation (σ'; 0.42 μeq L^{-1}) values of $nssSO_4^{2-}$ concentration in the layers, except those layers for which the values exceed $\bar{x}'+2\sigma$ are calculated. It is suggested that the layers in which the value of the $nssSO_4^{2-}$ concentration exceeds $\bar{x}'+2\sigma'$ (1.50 μeq L^{-1}) is affected by volcanic eruptions. However, as described in section 3.1, $nssSO_4^{2-}$ is produced by the oxidation of DMS (Nishio *et al.,* 2002; Suzuki *et al.,* 2005). In addition, a layer with high $nssSO_4^{2-}$ concentration resulting from volcanic eruptions has a low pH value (Suzuki *et al.,* 2005). By considering both the MSA^- and H^+ concentrations in the layers, we classify the volcanic signals into three grades. The layers whose $nssSO_4^{2-}$ concentration exceeds $\bar{x}'+2\sigma'$, H^+ concentration exceeds the average value (4.54 μeq L^{-1}), and MSA^- concentration doesn't indicate the peak were considered to preserve highly reliable volcanic signals and were classified as grade I. The layers whose $nssSO_4^{2-}$ concentration exceeds $\bar{x}'+2\sigma'$, and for which either the H^+ concentration exceeds the average value or the MSA^- concentration doesn't indicate the peak were classified as grade II. The layers whose H^+ concentration does not exceed the average value and whose MSA^- concentration indicate the peak were classified as grade III. Grade II and III layers were considered to be less affected by the volcanic eruptions and less reliable than grade I. Figure 6 shows the concentrations of H^+, MSA^-, and $nssSO_4^{2-}$ in the layers. It also shows a record of volcanic eruptions that occurred to the south of 20°N and that are thought to correspond to volcanic signals in the volcanic explosivity index (VEI) along with the degree of reliability of the layers (grade I, II, and III) in the right column. It can be seen that the ice core is affected by the volcanic eruptions of Pinatubo and Cerro Hudson in 1991, Agung in 1963, Tarawera in 1886, Krakatau in 1883 etc., except the three large volcanic eruptions used as time markers.

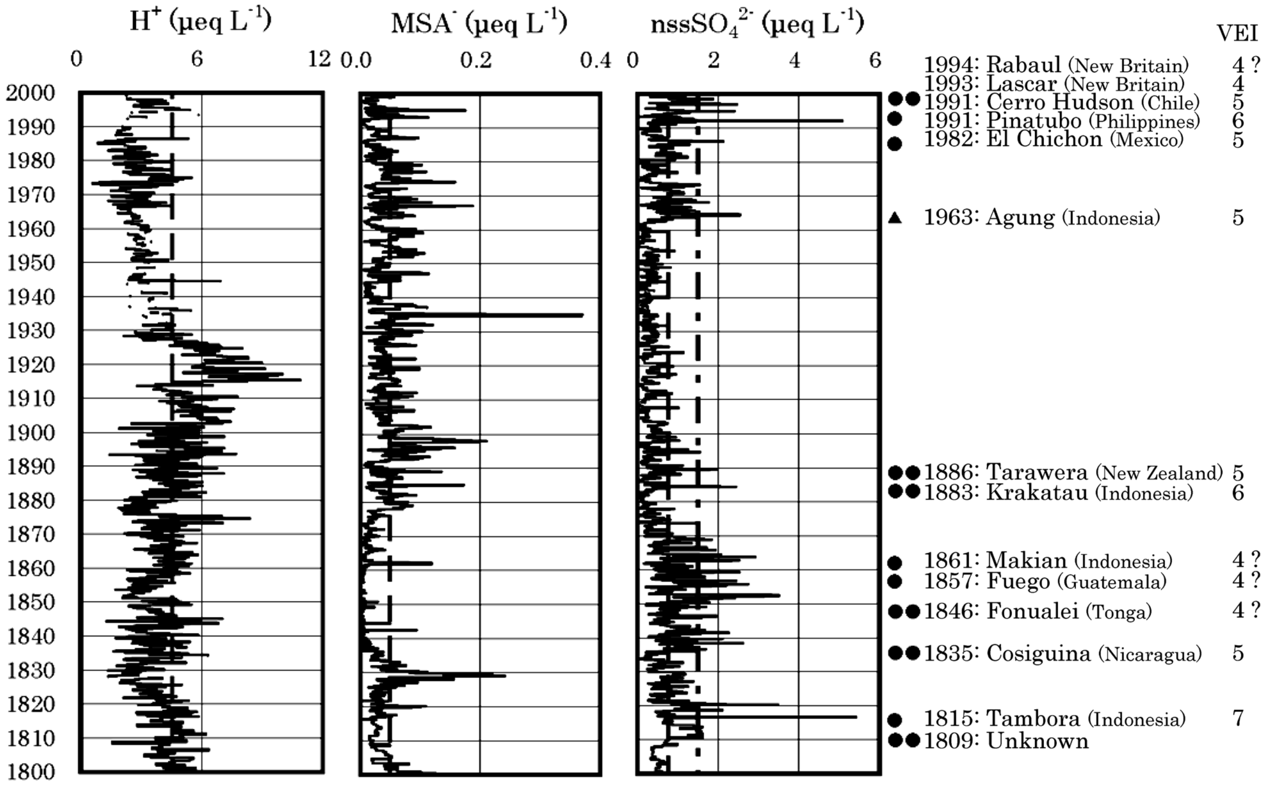

Fig. 6. Profiles of the concentrations of H^+, MSA^- and $nssSO_4^{2-}$ for last 200 years and the records of volcanic eruptions. The broken and the chain line in this figure indicate the average and $\bar{x}'+2\sigma'$, respectively. Symbols indicate degrees of reliabilities of volcanic eruptions (●●: grade I, ●: grade II, ▲: grade III). VEI values are compiled from the Global Volcanism Program (http: //www.volcano.si.edu/world/largeeruptions.cfm)

3.3.2 Relationship between climatic conditions and volcanic eruptions

Figure 7 shows the fluctuation in the $nssSO_4^{2-}$ concentration and $\delta^{18}O$ in the vicinity of the Tambora and Kuwae eruptions. $\delta^{18}O$ decreased significantly shortly after the $nssSO_4^{2-}$ peaks that are considered to include the effect of large volcanic signals. Sulfuric acid aerosols generated by volcanic eruptions remain suspended in the stratosphere for a period of a few years and are dispersed by solar radiation. These aerosols cause a decrease in the temperature on the earth's surface (Sigurdsson, 1990). It is suggested that large volcanic eruptions such as those of Tambora and Kuwae caused the decrease in the temperature at the YM85 site.

3.4 Surface mass balance

3.4.1 Surface mass balance in YM85

During 1453–1816 A.D., the surface mass balance at YM85 was significantly lower than those observed both prior to and subsequent to this period (Table 1). This is thought to be caused by two factors. One is the Little Ice Age (LIA) in the Northern Hemisphere. LIA is the cold period that lasted from the 14th to the 19th century in the Northern Hemisphere, and the ice cores obtained from the Southern Hemisphere also corroborate the occurrence of the LIA (Thompson *et al.,* 1986; Benoist *et al.,* 1982; Morgan *et al.,* 1985). It

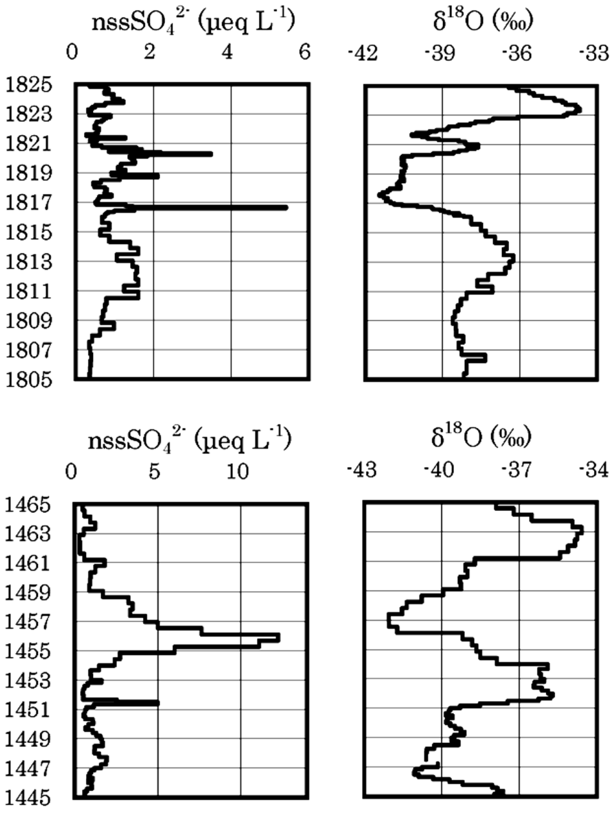

Fig. 7. Profiles of $nssSO_4^{2-}$ concentration and $\delta^{18}O$ in the vicinity of Tambora (above) and Kuwae (bottom) eruption. The broken and the chain line in this figure indicate the average and $\bar{x}'+2\sigma'$, respectively.

has been reported that the mean annual accumulation rate of snow at DT263, Antarctica is significantly low during the LIA. The other factor is a period during which there was a hiatus in snow deposition. The hiatus was observed at alternate regions separated by 10-km distances in the areas in the upper basin of the YM85 site (Furukawa *et al.,* 1996). Considering the annual average flow rate of the surface ice sheet at YM85 (as mentioned in section 2.1), the surface layer of the YM85 site is expected to have moved at the rate of 9 km year^{-1} for 363 years from 1454 to 1816 A.D. There is a possibility that the site where the next hiatus in snow deposition is likely to occur can be determined from the records preserved in the YM85 ice core. However, a sharp decrease in δ^{18}O during the period 1453–1816 A.D. is not observed in the ice core. The parameter that contains evidence of the hiatus in snow deposition preserved in the YM85 shallow ice core is as yet unknown. Further studies are necessary for explaining the significantly low surface mass balance during 1453–1816 A.D. at YM85.

Figure 8 shows the annual surface mass balance, which is calculated using the identified annual layers, for 182 years from 1816 to 1998 A.D. The average, maximum, and minimum values of the annual surface mass balance for this period are 141 mm w. e. year^{-1}, 336 mm w. e. year^{-1} in 1895, and 50 mm w. e. year^{-1} in 1937, respectively. The annual surface mass balance for the abovementioned period indicates a sharp decrease from the 19th to the 20th century, and the surface mass balance during the 19th century is higher than that during the 20th century. This pattern appears to be similar to that observed in the H72 ice core (Suzuki *et al.,* 2005), and hence, it could be expected that from the 19th to the 20th century, the annual surface mass balance showed a similar pattern throughout the area extending from the coastal region to approximately 200 km inland in East Dronning Maud Land.

3.4.2 Relationship between climatic conditions and surface mass balance

It has been previously reported that the fluctuation in δ^{18}O in the H72 ice core showed seasonal variations: the fluctuation was high in summer and low in winter (Suzuki *et al.,* 2005). However, the fluctuation in δ^{18}O in the YM85 shallow ice core did not show clear seasonal variations (Fig. 5). Using snow stake measurements, the average surface mass balance at the H72 site was calculated to be 307 mm w. e. year^{-1} for the period 1973–1998 (Nishio *et al.,* 2002); this value is much higher than the average surface mass balance at YM85 for the period 1816–1998 A.D. It is suggested that the value of δ^{18}O in the surface snow layer gradually changes because of isotopic fluctuations, and that the seasonal variations in δ^{18}O are not preserved in the YM85 core due to a low accumulation rate.

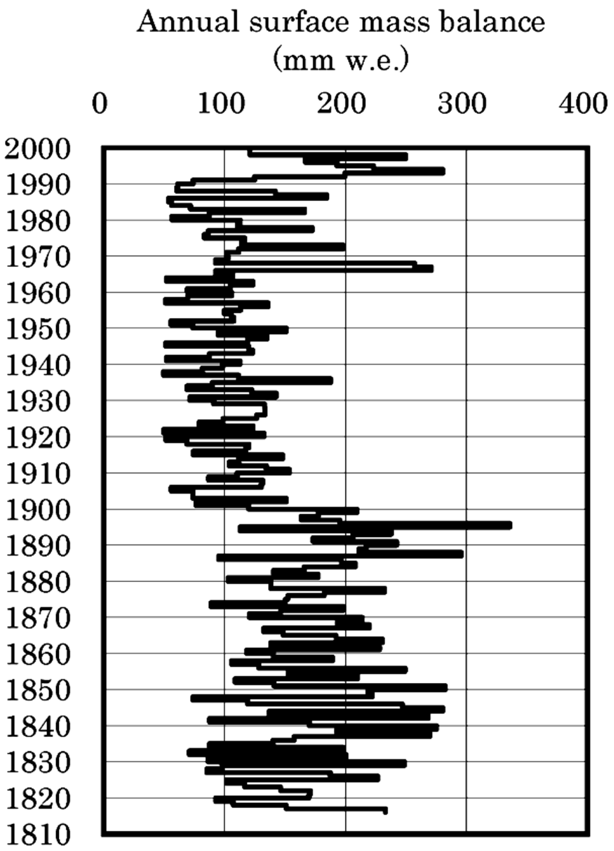

Fig. 8. Annual surface mass balance from 1816 to 1998 A.D.

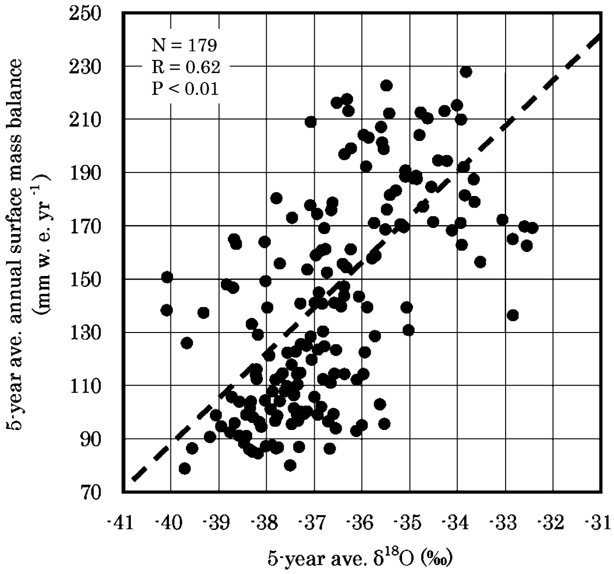

Fig. 9. The relationship between the 5-year-average of δ^{18}O and the 5-year-average of annual surface mass balance from 1816 to 1998 A.D. The broken line in this figure indicates a linear regression.

Figure 9 shows the relationship between the five-year average values of δ^{18}O and those of the annual surface mass balance for the period 1816–1998 A.D. If the five-year average of δ^{18}O is high, then the five-year average of annual surface mass balance is also high;

this indicates that the correlation coefficient (R=0.62) between the two parameters is positive. According to the relationship between the average temperature and the accumulation rate obtained for Greenland and Antarctica, the site with a high temperature tends to have a high accumulation rate (Satow *et al.,* 1999; Röthlisberger *et al.,* 2002). It is suggested that the period with a high temperature tends to show a high surface mass balance at YM85. The decrease in the concentration of Na^+ with an increase in the temperature, as mentioned in section 3.1, could be because of the occurrence of a large amount of snowfall. The decrease in the MSA^- and NO_3^- concentrations at the approximate depth ranges of 18–22 m w. e. and 36.5–38 m w. e. (Fig. 2) may be, to some extent, because of melting of the snow as a result of high temperature. However, the effect having the peculiar reaction processes by preserving into ice core, as described above in 3.1 section, can exceed the effect by the diluting since both MSA^- and NO_3^- concentrations show a significant decrease.

4. Conclusions

An ice core with a length of 77.71 m w. e. that was collected in January 2002 at YM85 in East Dronning Maud Land, Antarctica, by JARE-42 has been subjected to a detailed analysis. The layer at a depth of 77.71 m w. e in the ice core is dated at 1233 A.D using the tritium concentration and volcanic signal time markers and by annual layer counting, which involved counting of the number of annual layers using several ionic concentrations that were believed to show seasonal variations.

The ice core appears to have been affected by the volcanic eruptions of Pinatubo and Cerro Hudson in 1991, Agung in 1963, Tarawera in 1886, Krakatau in 1883, Tambora in 1815, Unknown in 1809, Kuwae in 1453, and Unknown in 1259. $\delta^{18}O$ shows a significant decrease shortly after large volcanic signals such as those from Tambora in 1815 and Kuwae in 1453, and hence, it is concluded that these large volcanic eruptions could have caused a decrease in the temperature at the YM85 site.

During 1453–1816 A.D., the surface mass balance at YM85 was significantly lower than that observed both prior to and subsequent to this period. This is thought to be caused by the LIA and a hiatus in snow deposition. However, this is debatable. Further studies are required to elucidate the reason for the low surface mass balance during this period.

An observation of the annual surface mass balance at YM85 for 182 years, from 1816 to 1998 A.D., indicates a sharp decrease from the 19th century to the 20th century; the surface mass balance during the 19th century is higher than that during the 20th century. This pattern appears to be similar to that observed in the H72 ice core, and hence, it could be expected that from the 19th century to the 20th century, the annual surface mass balance showed a similar pattern throughout the area extending from the coastal region to approximately 200 km inland in East Dronning Maud Land.

There is a clear relationship between the fluctuation in $\delta^{18}O$ and that in the MSA^- and NO_3^- concentrations in the YM85 ice core; hence, it is suggested that the processes of preservation of these ions is related to the temperature at which they are deposited. When comparing the fluctuation in $\delta^{18}O$ with that in the Na^+ concentration, a high $\delta^{18}O$ is found at low Na^+ concentrations. This is inferred to be because of the occurrence of a large amount of snowfall, which is in turn a consequence of a high surface mass balance.

It is necessary to assess the implications of the present study for the whole Antarctic ice sheet. Additionally, further research on the methods of analyzing ice cores is necessary to tackle unresolved problems.

Acknowledgments

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