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High-resolution 900 year volcanic and climatic record from the Vostok area, East Antarctica

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Abstract. Ion chromatography measurements of 1730 snow and firn samples obtained from three short cores and one pit in the Vostok station area, East Antarctica, allowed for the production of the combined volcanic record of the last 900 years (AD 1093–2010). The resolution of the record is 2–3 samples per accumulation year. In total, 24 volcanic events have been identified, including seven well-known low-latitude eruptions (Pinatubo 1991, Agung 1963, Krakatoa 1883, Tambora 1815, Huanaputina 1600, Kuwae 1452, El Chichon 1259) found in most of the polar ice cores. In comparison with three other East Antarctic volcanic records (South Pole, Plateau Remote and Dome C), the Vostok record contains more events within the last 900 years. The differences between the records may be explained by local glaciological conditions, volcanic detection methodology, and, probably, differences in atmospheric circulation patterns. The strongest volcanic signal (both in sulfate concentration and flux) was attributed to the AD 1452 Kuwae eruption, similar to the Plateau Remote and Talos Dome records. The average snow accumulation rate calculated between volcanic stratigraphic horizons for the period AD 1260–2010 is 20.9 mm H_2O. Positive (+13 %) anomalies of snow accumulation were found for AD 1661–1815 and AD 1992–2010, and negative (−12 %) for AD 1260–1601. We hypothesized that the changes in snow accumulation are associated with regional peculiarities in atmospheric transport.

1 Introduction

Polar ice cores contain information about atmospheric aerosols over continuous time intervals. Studies of these ice cores allowed reconstructing of global environmental change over the past several hundred thousand years (e.g., Petit et al., 1999; EPICA community members, 2004). High-resolution (years to decades) climate records from ice cores characterize environmental changes in more detail, although the mechanisms of these changes are not fully understood (Mosley-Thompson et al., 1993). Volcanic sulfate aerosol in the atmosphere is a possible factor in climatic fluctuations over relatively short time intervals. Explosive eruptions eject huge amounts of gas (mainly sulfur dioxide) and solid particles (usually silicate ash) into the stratosphere and have the capability of cooling the global climate by 0.2–0.7 °C for several years after the eruption (Zielinski, 2000; Cole-Dai et al., 1997). However, to obtain a complete picture of links between climate and volcanism, we need more high-resolution records of climate and volcanic changes, which can be easily obtained from ice cores. Volcanic events clearly fixed in these cores by horizons with higher acidity and electrical conductivity, by peaks of non-sea-salt sulphates, as well as particles of volcanic origin (tephra).

Climatic variations of the last millennium include both warm (the Medieval Warm Period of the 10–12th centuries and recent warming since the mid-20th century) and cold (the Little Ice Age of the 15–19th centuries) episodes and detailed study of ice cores taken at different sites allows a better understanding of the regional peculiarities of past

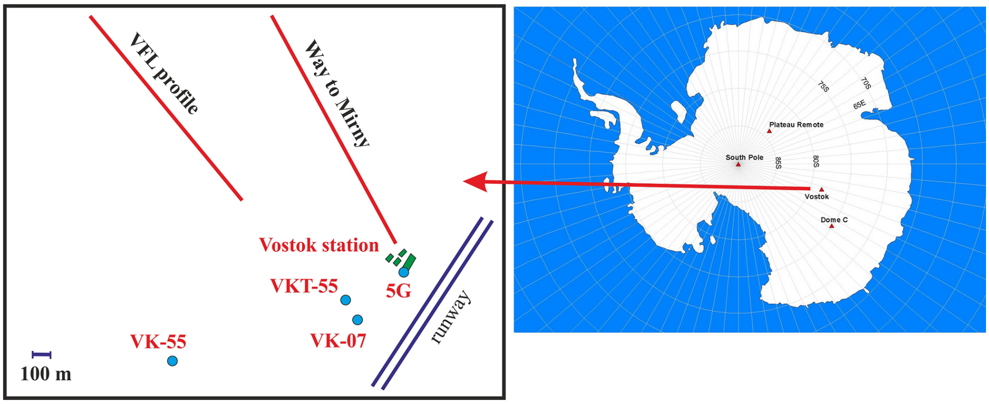

Fig. 1. Locations of the Vostok station and core sites used in this study (left side). Other Antarctica core sites mentioned in the text are also indicated (right side).

climates, including mechanisms of atmospheric circulation (Russel and McGregor, 2000). At present there are several high-resolution volcanic records from Antarctica covering at least the last millennium (e.g., Delmas et al., 1992, Cole-Dai et al., 2000, Castellano et al., 2005, Ferris et al., 2011, Jiang et al., 2012 and others). Here we try to extend this list and present a composed high-resolution volcanic and climatic record of the past 900 years from the Vostok station area, East Antarctica, the lowest snow accumulation site in Antarctica.

2 Data and methods

2.1 Ice and snow cores

In this study we analyze some snow/firn core sections from four sites located in the vicinity of the Russian Vostok station (Table 1, Fig. 1). The cores were taken in the course of Russian Antarctic Expeditions (RAE) during the summer seasons of 1990, 2007 and 2010. Directly in the field electrical conductivity was continuously measured over the cores and some sections containing suspected volcanic events were chosen for detailed chemical analysis. The cores were transported frozen in a pre-cleaned plastic package from Vostok station to the laboratory. Unfortunately we were not able to get a single long core from the summer snow surface, so we had to sew the available cores for a single combined record.

2.2 Sample preparation and analysis

Chemical analysis of the core was carried out in the Laboratory of Hydrochemistry and Chemistry of Atmosphere (Limnological Institute SB RAS, Irkutsk). Prior to the analysis the cores were stored in a frozen state (−20 °C). The outer parts of the cores (1.5–2.0 cm) were removed mechanically. For analysis we used only the central parts of the cores, which were cut with 2–3 cm steps in a laminar box under clean conditions. Then samples were melted and one part of a melted sample was used for measuring the pH at a temperature of 25 °C. Another part was filtered through the membrane filters (with 0.2 µm cell size) pre-washed by deionized water. Concentrations of major cations (Ca^{2+}, Mg^{2+}, Na^{+}, K^{+}) and anions (NO_3^-, Cl^-, SO_4^{2-}) were measured in the filtrate using an ICS-3000 (Dionex) ion chromatograph with manual sample introduction (by syringe). As a comparison, we also used standard solutions from the "Kanto Chemical Co" (Japan) and "Ultra Scientific" (USA). The accuracy of the used techniques is confirmed by involving the laboratory in the analysis of control samples in international programs "Global Atmospheric Chemistry" (GAW) of the WMO and "Acid Deposition Monitoring Network in East Asia" (EANET). International calibration reports are included in the WMO and EANET. Additionally, we have performed inter-laboratory comparison of ion measurements of 94 snow samples from East Antarctica with the Laboratory of Glaciology and Geophysics of the Environment – LGGE (Grenoble, France). Comparison analysis of SO_4^{2-} concentrations performed in two laboratories revealed the identity of their changes from sample to sample. About 75 % of the sulfate measurements obtained in two different laboratories had a discrepancy of less than 30 %, of which 32 % differed by less than 10 %. In total, about 1730 samples have been analyzed. It should be noted that the records from three ice cores (5G, VK-07 and VKT-55) are not continuous and have gaps, displayed in Table 1. Unfortunately, some segments of the cores (5G, VC-07) were either lost (i.e., as a result of storage or transportation) or completely used by other researchers (5G). Here we present only SO_4^{2-} measurement data. Based on snow accumulation data (see below), the sampling resolution for ion measurement is estimated to range from 1.8 (5G) to 3.0 (VK-55) samples per year.

2.3 Core dating

Snow depths were converted into a water equivalent by using the density data. Here we used a synthesis linear

Table 1. Description of snow/firn cores used in this study (for locations see also Fig. 1).

Site	Description	Location	Analyzed depth intervals, m
5G	> 3600 m-deep borehole (started in summer 1990 ∼ snow surface age)	Vostok station (78.465° S, 106.835° E)	15.50–36.00 (with total gap length 3.68 m)
VK-07	18 m borehole from the bottom of 2.2 m snow pit (RAE-52, January 2007)	∼ 300 m to the SW from 5G	3.80–19.30 (with total gap length 1.23 m)
VKT-55	Shallow borehole (RAE-55, summer 2010)	∼ 150 m from VK-07	2.10–4.48 (total gap length 0.80 m); 6.61–8.35
VK-55	2 m snow pit (RAE-55, summer 2010)	∼ 1.4 km to the SW from 5G	0.00–2.00

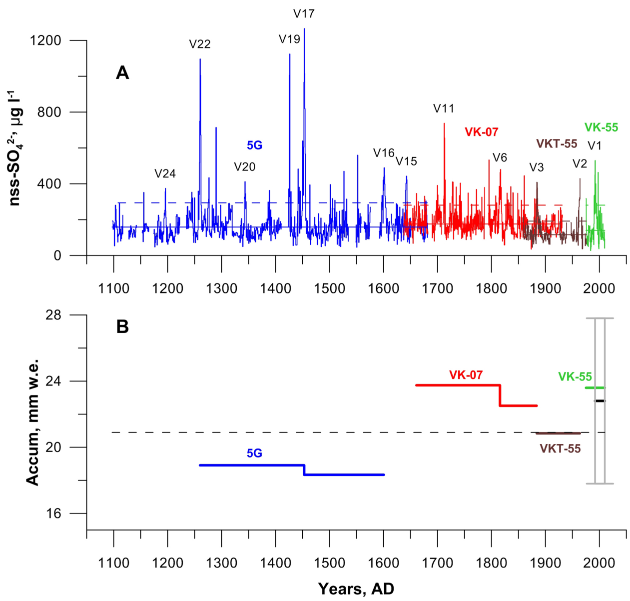

Fig. 2. (A) Nss-SO_4^{2-} volcanic Vostok record of the past 900 years. Each of used four cores is shown by a different color (5G, VK-07, VKT-55 and VK-55). Major volcanic eruptions are labeled (for more details see Table 3). Solid horizontal lines indicate a non-volcanic background and dashed lines show the detection threshold (explanations are in the text). **(B)** Accumulation rates calculated between the volcanic horizons in the studied cores. Black dashed line represents the long-term mean. Mean value of snow accumulation at Vostok station measured at the stack network for the period 1992–2010 is shown by a black horizontal line (vertical grey bars represent standard deviations).

density–depth relationship calculated in the Vostok area (Ekaykin et al., 2004) for the 5G, VK-07 and VKT-55 cores:

$$\rho = 7.46H + 350$$

and the individual linear relationship derived from direct density measurements for VK-55:

$$\rho = 46.75H + 362,$$

where ρ (kg m^{-3}) is the density at the depth H(m).

In total, eight volcanic stratigraphic markers have been used in this study (Table 2). These are the large eruptions of El Chichon (1259), Kuwae (1452), Huaynaputina (1600), Long Island (1660), Tambora (1815), Krakatoa (1883), Agung (1963) and Pinatubo (1991) previously revealed elsewhere in the Antarctica and Greenland ice cores (Legrand and Delmas, 1987; Moore et al., 1991; Delmas et al., 1992; Cole-Dai et al., 1997, 2000; Zhang et al., 2002 and others). The most important volcanic eruptions of the 19th century, Tambora (1815) and Krakatoa (1883), have been found in two cores at the Vostok station area, VK-07 and VFL-1 (Khodzher et al., 2011). Dating of the VK-55 core is quite accurate as based on a very prominent sulfate peak at the depth of 1.09 m attributed to the well-known eruption of Pinatubo (AD 1991). It has been taken into consideration that the time lag between the actual time of eruption and aerosol accumulation on the snow surface in Antarctica seems to be 1–2 years (Delmas et al., 1985). Average annual accumulation rates between two adjacent volcanic markers were calculated and intermediate horizons dated (Table 2). Below the lowest stratigraphic core marker the mean rate of overlaying horizons has been used. The age of the oldest sample (5G) has been dated to AD 1093. Thus the ion concentration profiles synthesized from four separate cores cover at least the last 900 years.

Three cores (5G, VK-07 and VKT-55) are linked by their overlapping parts (Fig. 2a, Table 2), which additionally confirm the accuracy of core dating. Firstly, there are two correlated nss-sulfate spikes at the depths of 17.03 m (5G, 443 µg L^{-1}) and 18.98 m (VK-07, 299 µg L^{-1}), dated at AD 1642 and 1644, respectively. Probably, these spikes would be attributed to the AD 1641 Deception Island (63° S, 61° W) eruption also previously found in the South Pole core (Delmas et al., 1992). Secondly, two spikes at depths of 15.63 m (5G, 304 µg L^{-1}) and 17.29 m (VK-07, 285 µg L^{-1}) were dated at AD 1678 and 1679, respectively. We can not accurately attribute these peaks to a particular eruption. Thirdly, prominent volcanic spikes of the Krakatoa eruption (1883) are seen in both the VK-07 (at a depth of 6.36 m, 368 µg L^{-1}) and VKT-55 (7.12 m, 409 µg L^{-1}) cores.

2.4 Identification of volcanic events

Non-volcanic sulfate comes to Antarctic snow from sea-salt spray, crustal erosion and atmospheric oxidation of

Table 2. Volcanic stratigraphic markers and their ages.

Core	Volcano (year of eruption, AD)	Depth in core, m	Depth in H_2O, m	Year in core[a], AD	Period covered	Mean accumulation, mm $H_2O\,yr^{-1}$
5G	Huaynaputina (1600)	18.60	8.70	1601		
					1453–1601	18.3
	Kuwae (1452)	23.93	11.41	1453		
					1260–1453	18.9
	El Chichon (1259)	30.52	15.06	1260		
					< 1260	19.6[b]
VK-07	Krakatoa (1883)	6.36	2.43	1884		
					1816–1884	22.5
	Tambora (1815)	10.08	3.96	1816		
					1661–1816	23.7
	Long Island (1660)	18.16	7.64	1661		
					<1661	23.4[b]
VKT-55	Agung (1963)	2.82	1.03	1964		
					1884–1964	20.8
	Krakatoa (1883)	7.12	2.70	1884		
					< 1884	20.8[b]
VK-55					1992–2010	23.6
	Pinatubo (1991)	1.09	0.43	1992		
					< 1993	23.6[b]

[a] Assigned years to the volcanic signals (taking into consideration a one-year lag between a volcanic eruption and sulfate deposition). [b] Extrapolated values.

biogenic dimethylsulfide (DMS). The contribution of sea-salt SO_4^{2-} to the total sulfate budget at Vostok, calculated from the SO_4^{2-} / Na^+ ratio in bulk sea water, is less than 7 %. The main source of non-volcanic sulfate in Antarctic snow is biogenic DMS. In order to estimate the background content of non-sea-salt (nss) SO_4^{2-} we used the approach previously realized in other studies (e.g., Cole-Dai et al., 2000). As the studied cores cover the different time intervals with quite expectable natural changes in sulfate content here we calculated the background values for each core. The values were calculated as the average of the nss-SO_4^{2-} concentration after manual removal of some of the highest (presumably volcanic origin) concentrations in the cores. Respectively, the non-volcanic background for the 5G core was estimated as $159 \pm 67\,\mu g\,L^{-1}$, for VK-07 as $176 \pm 52\,\mu g\,L^{-1}$, for VKT-55 as $116 \pm 38\,\mu g\,L^{-1}$ and for VK-55 as $140 \pm 70\,\mu g\,L^{-1}$. The average background sulfate value for all cores is $160 \pm 62\,\mu g\,L^{-1}$.

Following Cole-Dai et al. (1997, 2000) we also calculated the volcanic threshold as the average nss-sulfate concentration plus 2σ. The thresholds for each core were used here as a criterion for distinguishing explosive volcanic events from the background ($294\,\mu g\,L^{-1}$ for 5G, $281\,\mu g\,L^{-1}$ for VK-07, $192\,\mu g\,L^{-1}$ for VKT-55 and $281\,\mu g\,L^{-1}$ for VK-55). Moreover, an excess nss-SO_4^{2-}concentration above the threshold for not less than two successive samples was an additional criterion for detecting a volcanic event.

For each volcanic event its duration (in years) and sulfate flux (in $kg\,km^{-2}$) were calculated. As is known, in areas dominated by dry sulfate deposition (low accumulation areas) the effect of the variability of snow accumulation rates could be smoothed by considering the sulfate flux (Legrand and Delmas, 1987; Castellano et al., 2005). The volcanic sulfate flux was obtained for each sample within the volcanic event by multiplying the net sulfate concentration (the difference between nss-sulfate concentration and background) by the sample length in water equivalent. The volcanic sulfate flux of an event was obtained as the cumulative sum of sulfate fluxes of separate samples.

3 Result and discussion

3.1 900 year Vostok volcanic record

The non-sea-salt SO_4^{2-} concentration profile demonstrates a strong variability over the past 900 years, with an average of $171 \pm 101\,\mu g\,L^{-1}$ (Fig. 2a). Using the criteria mentioned above we detected 24 volcanic eruptions from the combined Vostok sulfate record (VR) between AD 1093 and 2010. The identified volcanic events (V1 to V24) are listed in Table 3 and marked in Fig. 2a. Ten of them have been attributed to historically documented eruptions (e.g., by Simkin and Siebert, 1994). These include the well-known eruptions detected overall in Antarctic ice cores covering the last millennium: V2-Agung (1963); V3-Krakatoa (1883); V6-Tambora (1815); V7-Unknown (1809); V16-Huaynaputina (1600); V17-Kuwae (1453) and V22-El Chichon (1259). The largest number of volcanic eruptions occurred in the 19th and 18th (5 events) centuries. However, the most powerful eruptions, in accordance with the VR, occurred in the 13th and 15th centuries. The largest event marked both by sulfate concentration ($1266\,\mu g\,L^{-1}$) and flux ($61.4\,kg\,km^{-2}$ or almost three times more than after the Tambora eruption) is the 1452 eruption of Kuwae, Vanuatu, South Pacific (17° S, 168° E). The second large event of the VR is the V22 AD 1260 event (5G core). This eruption has been tentatively attributed to El Chichon (17° N, 93° W) in Mexico (Palais et al., 1990; Cole-Dai et al., 2000). Langway et al. (1988) used this signal as a bipolar stratigraphic ice core marker. The duration of the events varies from 0.7 (V8) to 8.8 (V16) years. Tambora-normalized values ($F/F_{tambora}$) in the VR vary from 0.13 to 2.88, although most of them are lower than 1.0 (Table 3). The exceptions are three events, AD 1452 Kuwae (2.88), AD 1259 El Chichon (2.13) and AD 1600 Huaynaputina (1.29). It means that the Tambora eruption (AD 1815) was the strongest event just over the last 400 years.

We compared the Vostok volcanic record with those from inland areas of East Antarctica, such as the South Pole (SP, Delmas et al., 1992), Plateau Remote (PR, Cole-Dai et al., 2000) and Dome C (DC, Castellano et al., 2005). Locations of the cored sites are shown in Fig. 1 and correlations of volcanic events from four sites in Table 4 and in Fig. 3. In contrast to other records the VR contains almost two times more volcanic events. Only 16 events of the VR were found in at least one of the mentioned records (PR, SP and DC) for the comparable period. On the other hand, eight VR events (V8–V11, V14–V15, V18–V19) were not found in other cores. Some of the events are characterized by very high sulfate concentrations, for example, very prominent peaks V11 ($738\,\mu g\,L^{-1}$, AD 1712) and V19 ($1124\,\mu g\,L^{-1}$, AD 1426). VR also contains more large to moderate events grouped near the AD 1452 Kuwae eruption (V17–V19). The largest number of VR events not revealed in any other record is within the period between AD 1700 and 1808. Earlier, Liya et al. (2006) also revealed several abnormal peaks of SO_4^{2-} of volcanic origin within the AD 1460–1800 period in the DT263 East Antarctic ice core. However, they could not identify the individual volcanic events due to lower accumulation at that time.

In general, a larger number of eruptions in the VR seem to be due to higher sulfate concentrations at Vostok. For example, sulfate spikes of seven well-known volcanic events in the VR (V2–V3, V6–V7, V16–V17 and V22) are higher than those in the other three East Antarctic cores, on average 1.8 times. It should be noted that sulfate excess was observed not only for the Tambora layer (V6; ratios are from 0.6 to 1.0). On the other hand, sulfate background in the VR is also increased (e.g., 160 vs. $95\,\mu g\,L^{-1}$ in the PR). A possible explanation for the higher sulfate level is the lower accumulation rate at Vostok than at other sites (22 vs. 27–80 mm $H_2O\,yr^{-1}$) coupled with primary (> 80 %) dry sulfate deposition (with a very small dilution effect). Legrand and Delmas (1987) reported a relation between enhanced volcanic fallout from the stratosphere in central Antarctic areas and a dry deposition mechanism for the Tambora and Agung eruptions. They also showed that some volcanic SO_4^{2-} spikes at Vostok are much higher and prominent than at other East Antarctic sites (e.g., 1.5–3.0 times in comparison with Dome C). Besides, it should be noted that there are obvious differences in methodologies for defining volcanic thresholds (e.g., between two low-accumulation sites, Vostok and Dome C). The number of revealed volcanic events depends closely on the selected detection method, although it mostly concerns only minor volcanic eruptions.

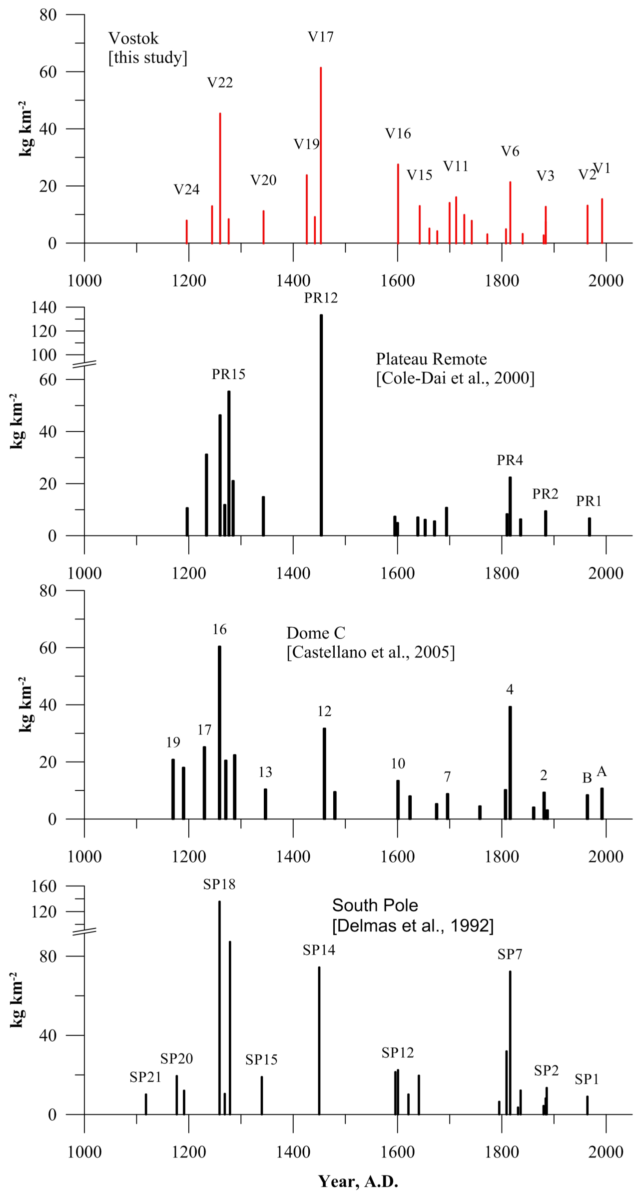

Fig. 3. Comparison of sulfate fluxes of volcanic eruptions recorded in four East Antarctic cores. Explanations are in the text.

Table 3. Volcanic events for the last 900 years in the Vostok area (combined).

Core	Eruption (year)	Event	Depth in core, m	Peak nss-sulfate, $\mu g\,L^{-1}$	Year, AD	Duration, years	Volcanic flux (F), kg km^{-2}	$F/F_{tambora}$
VK-55	Pinatubo (1991)	V1	1.09	530	1992	4.5	15.4	0.72
VKT-55	Agung (1963)	V2	2.82	429	1964	3.2	13.1	0.62
VKT-55	Krakatoa (1883)	V3	7.12	409	1884	3.6	12.7	0.60
VK-07	Krakatoa (1883)	V3	6.36	368	1884	3.7	7.5	0.35
VK-07	unknown	V4	6.58	317	1880	1.8	2.7	0.13
VK-07	unknown	V5	8.78	351	1840	1.1	3.2	0.15
VK-07	Tambora (1815)	V6	10.08	464	1816	5.7	21.3	1.00
VK-07	Unknown (1809)	V7	10.53	333	1808	1.8	4.9	0.23
VK-07	unknown	V8	12.47	368	1772	0.7	3.1	0.15
VK-07	unknown	V9	14.06	404	1742	2.3	7.8	0.37
VK-07	unknown	V10	14.81	411	1728	2.7	9.9	0.46
VK-07	unknown	V11	15.61	738	1712	3.2	16.1	0.75
VK-07	unknown	V12	16.25	391	1700	8.1	14.1	0.66
VK-07	unknown	V13	17.43	331	1676	2.3	4.2	0.20
VK-07	Long Island (1660)	V14	18.16	346	1661	3.3	5.1	0.24
5G	Parker+Deception (1641)	V15	17.03	443	1642	2.6	13.0	0.61
5G	Huaynaputina (1600)	V16	18.60	489	1601	8.8	27.5	1.29
5G	Kuwae (1452)	V17	23.93	1266	1453	8.0	61.4	2.88
5G	unknown	V18	24.33	483	1442	2.3	9.2	0.43
5G	unknown	V19	24.87	1124	1426	2.3	23.8	1.11
5G	unknown	V20	27.71	412	1343	2.9	11.2	0.53
5G	unknown	V21	29.97	434	1276	2.7	8.4	0.39
5G	El Chichon (1259)	V22	30.52	1097	1260	5.9	45.4	2.13
5G	unknown	V23	31.04	353	1245	7.0	12.9	0.61
5G	unknown	V24	32.68	374	1196	4.2	7.9	0.37

The twin peaks of the volcanic eruptions of Krakatoa, SP3 (1883) and Tarawera, SP2 (New Zealand, 1886) were found in the SP record (Delmas et al., 1992). Although we also see the peak dated as AD 1886 in the VR, it does not rise above the threshold, unlike the Krakatoa spike (V3). At the same time, events V6 (AD 1816) and V7 (AD 1808) of the VR are also separately presented in other cores (5 and 4 at Dome C, PR5 and PR4 at Plateau Remote and SP8 and SP7 at the South Pole, respectively). It suggests that mixing of snow layers in Vostok area is estimated to be not more than 3–8 years, at least in the 19th century.

The sulfate flux of the Pinatubo (AD 1991) eruption at the Vostok site was about 1.4 times more than at Dome C. Higher flux values at Vostok are observed for the Agung (1963) and Huaynaputina (1600) eruptions (Fig. 3). However, the Tambora eruption (1815) flux in the VR is similar to that in PR and less than in DC and SP. The flux from the Kuwae eruption (1452) in the VR is less significant than in PR and SP and more than in DC. Moreover, the VR flux of Kuwae is about 1.4 times more than AD 1259, similar to PR, although in the SP (Delmas et al., 1992) and DC (Castellano et al., 2005) records the Kuwae volcanic flux is significantly less than that of the AD 1259 eruption. Cole-Dai et al. (2000) explained the lower sulfate flux of the "1259 event" in the PR by a possible loss of the volcanic mass due to snow drift and redistribution (impact of the local glaciology). On the other hand, as in the VR, the Kuwae eruption in the Talos Dome ice core is the largest volcanic signal of the last 800 years (Stenni et al., 2002). The mentioned differences in volcanic fluxes between volcanic records are probably caused by both spatial differences in sulfate fallout and local accumulation conditions (e.g., wind redistribution of snow).

While the major volcanic events have been identified in the studied cores, some caution should be kept in mind when interpreting the data of the 5G core. It should be noted that the core was obtained by thermal drilling (Vassilev et al., 2007), which could produce possible contamination due to percolation of the melt water into porous firn. Despite the applied decontamination procedures (e.g., removing the outer parts of the core, see above), we can assume that some suspicious events, such as V19, with an nss-SO_4^{2-} flux comparable with the Tambora eruption, not seen however in any other record, may be spurious. However, we need more information to confirm this unequivocally.

3.2 Snow accumulation rates

Snow accumulation rates calculated between some volcanic horizons during AD 1260–2010 vary from 18.3 to 23.7 mm w.e., with a long-term mean of 20.9 mm (Table 2, Fig. 2b). During AD 1260–1601 mean accumulation rates were lower than the long-term mean by 10–12 %, while between AD 1661–1815 and AD 1992–2010 they were higher by 13 %. The period AD 1884–1964 is characterized by an

Table 4. Correlation of volcanic events of the Vostok record with other cores in East Antarctica.

	Vostok (this study)		Plateau Remote (Cole-Dai et al., 2000)		Dome C (Castellano et al., 2005)		South Pole (Delmas et al., 1992)	
Eruption (year)	Event	Year, AD	Event	Year, AD	Event	Year, AD	Event	Year, AD
Pinatubo (1991)	V1	1992			A	1992		
Agung (1963)	V2	1964	PR1	1968	B	1964	SP1	1964
Krakatoa (1883)	V3	1884	PR2	1884	1	1887	SP3	1884
unknown	V4	1880			2	1881		
unknown	V5	1840	PR3	1836			SP5	1836
Tambora (1815)	V6	1816	PR4	1816	4	1816	SP7	1816
Unknown (1809)	V7	1808	PR5	1810	5	1807	SP8	1809
unknown	V8	1772						
unknown	V9	1742						
unknown	V10	1728						
unknown	V11	1712						
unknown	V12	1700	PR6	1694				
unknown	V13	1676	PR7	1671	8	1675		
Long Island (1660)	V14	1661						
Parker+Deception (1641)	V15	1642						
Huaynaputina (1600)	V16	1601	PR10	1600	10	1601	SP12	1601
Kuwae (1452)	V17	1453	PR12	1454	12	1460	SP14	1450
unknown	V18	1442						
unknown	V19	1426						
unknown	V20	1343					SP15	1340
unknown	V21	1276	PR15	1277			SP16	1279
El Chichon (1259)	V22	1260	PR17	1260	16	1259	SP18	1259
unknown	V23	1245			17	1230		
unknown	V24	1196	PR19	1197	18	1190	SP19	1191

accumulation rate close to the AD 1260–2010 mean. Although the accuracy of accumulation rate calculations by volcanic horizons in low-accumulation areas depends strongly on the local glaciological conditions, we assume that the mentioned temporal variations in accumulation are, likely, related to climate changes. Firstly, the spatial and temporal variability of accumulation can be smoothed by averaging over longer (e.g., decadal) time periods. Petit et al. (1982) have shown that variability of the accumulation rate in the Dome C area becomes small on a 10-year scale. Despite the high interannual variability of measured accumulation at the Vostok station stack network (1.5 km to the north of the station; two perpendicular profiles, each being 1 km long; the total number of stakes is 79; the distance between adjacent stakes is 25 m) (Ekaykin, 2003) during AD 1992–2010 (22.8 ± 5.0 mm, Fig. 2), due to a local glaciology its mean value corresponds well to that calculated between the Pinatubo horizon and the snow surface (23.6 mm). Secondly, our reconstruction of the accumulation is confirmed by independent estimates. For example, Ekaykin et al. (2004), based on observations in some deep pits in the Vostok station area, concluded that the accumulation had a slight increase during the last 200 years. In accordance with our estimates there is also a positive accumulation trend between AD 1816 and 2010 (+2.7 mm or 0.014 mm yr^{-1}).

It is considered that during the cold intervals the snow accumulation rates were decreased. Many Antarctic proxies of the period approximately from AD 1400 to 1900 show cooler climate conditions related to the Little Ice Age (LIA) cold stage (e.g., Li et al., 2009; Bertler et al., 2011). Orsi et al. (2012) estimated that the temperature in West Antarctica during AD 1400–1800 was on average 0.52 ± 0.28 °C colder than the last 100 year average. Simms et al. (2012) found that Neoglacial advance in the South Shetland Islands (northern Antarctic Peninsula) occurred between AD 1500 and 1700. In general, there are sufficiently convincing data on accumulation changes in East Antarctica during the LIA time. Li et al. (2009) revealed sharply reduced snow accumulation rates between AD 1450 and 1850 in the core from Princess Elizabeth Land. A reduction in the accumulation rate during the time period of AD 1500 to 1900 has been, recently, found in the 1830 year South Pole record (Ferris et al., 2011). However, Mosley-Thompson et al. (1993) reported that net accumulation at the South Pole and in central Greenland was below the long-term mean from AD 1600 to 1700 and above from AD 1700 onward, while at Siple prior to AD 1700 accumulation was above average. The authors believed that such spatial differences result from minor shifts in preferred locations for large-scale circulation features. Decreased snow accumulation has been revealed in the ice core from Dronning Maud Land for the period from 1452 to 1641 (Karlöf et al., 2000). Sharply lower accumulation rates at the period AD 1460–1800 were determined at the DT263 site, East Antarctica, by Liya et al. (2006).

We can assume that the positive anomaly of snow accumulation at Vostok during the 17th and 18th centuries may reflect regional peculiarities in atmospheric transport during the final phase of the LIA. Udisti et al. (2004) studied relative changes in snow accumulation at the Vostok and Dome C sites during the last 45 kyr and concluded that the changes could be related to variations in regional atmospheric circulation. Recent measurements of snow accumulation along the ice flow lines passing through Lake Vostok have shown that this area is characterized by a strong latitudinal gradient of the snow accumulation rate (Ekaykin et al., 2012). It is likely that the Vostok site is very sensitive to changes in atmospheric circulation. An alternative interpretation of increased accumulation may be snow/ice transfer from a higher accumulation site along the flow line from Dome B to Vostok (VFL). Although the averaged rates along the VFL have very little change (21–24 mm H_2O), there is a sharp spatial variability in accumulation rates along the VFL (Ekaykin et al., 2012). Most probably these anomalies are related to a snow re-deposition as a result of the wind interplay with the glacier surface relief. However, over the long-term periods this factor seems to be negligible. Nevertheless, these assumptions are largely speculative and require further proof.

4 Conclusion

Four snow and firn cores taken in the Vostok site area have been analyzed for chemical composition (major ions). Analysis of the SO_4^{2-} profile allowed for the determination of 24 volcanic events within the last 900 years (AD 1093–2010). They include both well-known major eruptions of the last millennium (e.g., Pinatubo 1991, Agung 1963, Krakatoa 1883, Tambora 1815, Huaynaputina 1600, and Kuwae 1452) found elsewhere in Antarctica and those of more moderate intensity. The largest (both in sulfate concentration and flux) volcanic event in the Vostok record was the AD 1452 Kuwae eruption, similar to some other Antarctic records (e.g., Plateau Remote, Talos Dome). The sulfate volcanic record from Vostok contains more events compared with those from other sites (e.g., the South Pole, Plateau Remote, Dome C), particularly between AD 1600 and 1800. The differences between the records are related to higher (1.8 times) concentrations of sulfate spikes due to lower accumulation rates at Vostok, dry deposition with a minimal dilution effect, differences in volcanic detection approach, and, probably, differences in atmospheric circulation patterns coupled with local glaciological effects (e.g., wind redistribution of snow).

The average snow accumulation rate from AD 1260 to 2010 was calculated as 20.9 mm H_2O. Increased snow accumulation rates were determined from AD 1661 to 1815 and AD 1992 to 2010 (13 % above the long-term mean), and decreased from AD 1260 to 1601 (10–12 % below the long-term mean). Probably, the positive accumulation anomaly during the 17th and 18th centuries may reflect regional peculiarities in atmospheric transport during the final phase of the LIA.

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