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Meteorological regime of central Antarctica and its role in the formation of isotope composition of snow thickness

Alexey Ekaykin

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presented by

Alexey EKAYKIN

Meteorological regime of central Antarctica and its role in the formation of isotope composition of snow thickness

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**Laboratoire de Glaciologie et Géophysique de l'Environnement
(UMR CNRS-UJF, Grenoble, France)**

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Faculty of Geography and Geoecology

Board of the jury:

M-me Valerie Masson-Delmotte	President of jury
Mr. Jean Robert Petit	Director of thesis
Mr. Pavel Arapov	Co-director of thesis
Mr. Vladimir Lipenkov	Reviewer

Foreword

This work was carried out in the frame of a "co-tutelle" (meaning in French "joint supervision of a thesis") project between Université Joseph Fourier (UJF, Grenoble, France) and St. Petersburg State University (SPSU, St. Petersburg, Russia). In France, my host laboratory was LGGE (Laboratoire de Glaciologie et Géophysique de l'Environnement, Grenoble) and at the same time I was employed as a research scientist at Arctic and Antarctic Research Institute (AARI) in St. Petersburg. The idea of this study was born in the head of Dr. Vladimir Lipenkov, leading scientist of the Department of Polar Regions Geography of AARI. Dr. Jean Robert Petit (LGGE) and Prof. Pavel P. Arapov (SPSU) kindly agreed to be my supervisors, and since 1998 they directed my scientific activity.

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I am so much grateful to all my colleagues from the Department of Polar Regions Geography of AARI headed by Lev Savatyugin, especially to Ira Kuzmina and Alla Preobrazhenskaya;

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I would like to dedicate this work to my family – to my mother Tatiana, to Natalia, and to my sister Lena, in appreciation of their love and care...

Régime météorologique de l'Antarctique central et son rôle dans la composition isotopique de la neige

Résumé

L'objectif principal du travail présenté ici est d'une part la détermination de la relation entre la température locale de l'air et la composition isotopique de la neige à la station de Vostok (Antarctique de l'Est) et d'autre part la reconstruction de variabilité climatique dans cette région au cours des 200 dernières années.

On a produit des données expérimentales concernant la composition isotopique et l'accumulation de la neige à partir de 8 puits creusés à Vostok dont 2 couvrent les 200 dernières années et qui ont été réalisés pendant les saisons d'été 1998/99, 1999/2000 et 2001/02. Le régime météorologique est documenté par les observations instrumentales et les radiosondages de l'atmosphère (depuis Décembre 1957) archivés à l'Institut Recherche Arctique et Antarctique (St. Pétersbourg, Russie). Nous avons aussi utilisé les données d'accumulation de la neige venant du réseau des balises installé à proximité de la station de Vostok en 1970, documenté la distribution spatiale de la teneur isotopique de la neige de surface, mesuré le nivellement de la surface, et étudié la variation de la composition isotopique des précipitations et celle de la neige transportée par vent, qui ont été collectés de Décembre 1999 à Décembre 2000.

En comparant les données météorologiques et les données de l'accumulation de la neige, nous avons établi que la plupart des précipitations au centre de l'Antarctique était formée pendant des conditions de "ciel clair" (diamond dust). Par ailleurs les données de sondages aérologiques suggèrent que la température moyenne de condensation à Vostok est très proche de la température de la limite haute de la couche d'inversion.

Pour un site donné, on a mis en évidence que la variabilité temporelle de la composition isotopique de la neige est influencé par le relief de la surface de neige (microrelief, "mésodunes" et probablement mégadunes). Cet effet diminue considérablement le rapport "signal sur bruit".

On met en évidence que les variations saisonnières des teneurs isotopiques des précipitations suivent bien celles de la température locale. L'excès en deutérium varie au cours de l'année et il est, quant à lui, influencé par la source d'évaporation.

Pour les 40 dernières années et pour l'échelle de temps multi-décennale, une relation linéaire significative est observée entre la moyenne de la composition isotopique de la neige et la température de l'air près de la surface. Mais pour des échelles de temps plus courtes (10 ans) les variations isotopiques apparaissent dominées par les conditions de la source.

Pendant les deux derniers siècles, l'accumulation et la composition isotopique de la neige présentent des fluctuations avec une période de 50 ans environ, qui ont des similarités avec l'indice de l'Oscillation Décennale du Pacifique. Si cela est vérifié, cette relation impliquerait une téléconnection climatique de l'Antarctique centrale avec Pacifique tropical. Par ailleurs, l'excès en deutérium suggère un changement brusque vers 1963 dans la circulation atmosphérique qui alimente la région de Vostok en l'humidité.

Метеорологический режим центральной Антарктиды и его роль в формировании изотопного состава снежной толщи

Абстракт

Главными целями работы являются, во-первых, определение связи между приземной температурой воздуха и изотопным составом снега на станции Восток (Восточная Антарктида) и, во-вторых, восстановление изменчивости климата в этом районе за последние 200 лет.

Экспериментальная база настоящего исследования включает данные об изотопном составе и скорости накопления снега по 6-ти мелким и 2-м глубоким шурфам, вскрытым в районе Востока в летние сезоны 1998/99, 1999/2000 и 2001/02 гг. Метеорологический режим задокументирован путем инструментальных метеорологических и аэрологических наблюдений (с декабря 1957 г.), результаты которых хранятся в архиве Арктического и антарктического научно-исследовательского института (Санкт-Петербург, Россия). Также использованы данные о накоплении снега на речном снегомерном полигоне, установленном в окрестностях станции в 1970 г., данные о пространственном распределении изотопного состава поверхностного слоя снега, данные о высоте снежной поверхности и об изотопном составе снежных осадков и переметенного снега, образцы которых собирались в период с декабря 1999 по декабрь 2000 г.

Сопоставление метеорологических данных и результатов снегомерных наблюдений позволило установить, что большая часть осадков в центральной Антарктиде формируется в условиях ясного неба (т.н. "ледяные иглы"). Аэрологические данные предполагают, что средняя температура конденсации на Востоке очень близка температуре на верхней границе слоя приземной инверсии.

Установлено, что временная изменчивость изотопного состава снега в отдельно взятой точке определяется, в первую очередь, влиянием рельефа снежной поверхности (микрорельеф, "мезо-дюны" и, возможно, мега-дюны), что существенно понижает вклад климатического сигнала в общую изменчивость изотопного состава.

Сезонные вариации изотопного состава осадков тесно коррелируют с внутригодовой изменчивостью температуры воздуха. Вместе с тем, сезонная изменчивость эксцесса дейтерия осадков предполагает существенное влияние условий в источнике влаги на формирование изотопного состава снега в центральной Антарктиде.

Наблюдается значимая линейная корреляция между сводным рядом изотопного состава снега и приземной температурой воздуха за последние 40 лет. Вместе с тем, короткопериодные (~10 лет) колебания изотопного состава, видимо, связаны с влиянием источника влаги.

За последние 200 лет колебания изотопного состава и накопления снега обнаруживают цикл с периодом около 50 лет, который хорошо коррелирует с индексом Тихоокеанского Многолетнего Колебания. Этот факт предполагает наличие дальней связи между климатами центральной Антарктиды и тропической части Тихого океана. В то же время, данные об эксцессе дейтерия свидетельствуют о резком изменении атмосферной циркуляции, снабжающей влагой район Востока, около 1963 г.

Meteorological regime of central Antarctica and its role in the formation of isotope composition of the snow

Abstract

The main goals of the present study are, first, determination of the relationship between local air temperature and snow isotope composition at Vostok Station (East Antarctica) and, second, reconstruction of climatic variability in this area over the past 200 years.

Experimental basis of the study includes data on isotope composition and accumulation rate of snow from 6 shallow and 2 deep snow pits dug in the Vostok's vicinities during summer seasons 1998/99, 1999/2000 and 2001/02. Meteorological regime is documented by the results of instrumental meteorological and balloon-sounding observations being carried out at Vostok since December 1957 and archived in Arctic and Antarctic Research Institute (St. Petersburg, Russia). We also used the data on snow accumulation at the stake network constructed near the station in 1970, documented spatial distribution of surface snow isotope composition and studied variations of isotope composition in precipitating and blowing snow collected from December 1999 to December 2000.

By comparing meteorological and snow accumulation data, it has been established that most of precipitation in central Antarctica form under clear-sky conditions (diamond dust). Balloon-sounding data suggests that overall condensation temperature at Vostok does not significantly differ from the temperature at the top of the inversion layer.

Temporal variability of isotope composition in a single point is dominated by influence of snow relief (micro-relief, "meso-dunes" and possibly mega-dunes), which substantially diminishes the signal-to-noise ratio.

Seasonal variations of isotope composition of the precipitation closely follow those of the local temperature, though the influence of moisture source is also evident from the intra-annual changes of deuterium excess content.

Significant linear relationship is observed for the last 40 years between stacked series of snow isotope composition from pits and surface air temperature for the multi-decadal time-scale, while shorter (10-year) isotope variability is likely dominated by source conditions.

During the past two centuries both snow accumulation and isotope composition of snow display oscillations with an apparent period of about 50 years. Relationship between these variations and the Pacific Decadal Oscillation index is suggested. If true, this would imply a climatic teleconnection of central East Antarctica with tropical Pacific. Finally, deuterium excess data suggest a sharp change in atmospheric circulation supplying Vostok area with moisture around 1963.

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INTRODUCTION

The necessity of comprehensive study of Antarctic environment is primarily related to the existence of the only Earth's continental glaciation playing an important role in the formation of climate of the whole planet. The uniqueness of Antarctic environment is secured in the special political status of this mainland, which represents an international area for implementing various scientific researches. One of the most productive and fruitful areas of activity is the study of physical and chemical properties of the ice deposits representing a unique natural archive of paleo-climatic information. In particular, completing of long-term project of deep ice coring at Russian Vostok Station allowed detailed investigating of the climate of this region over the last 420 kyrs [Petit et al., 1999]. Since recently, more and more attention is attracted to the subglacier water bodies known as "subglacial lakes". The study of accretion ice formed from the water of the largest of them, Lake Vostok, has already changed the ideas about the possible limits of life on Earth [Bulat et al., 2003; Lukin et al., 2003]. Owing to their extreme environmental conditions, the subglacial lakes are considered as terrestrial analogues of the oceans existing beneath ice sheets that cover some of the moons of big planets of Solar system.

The first human footstep on the Antarctic ground was only made about 80 years after the sixth continent had been discovered by the Russian marine expedition under the command of F.F. Bellingshausen and M.P. Lazarev on the 28th of January 1820. In 1899 the Norwegian group headed by C. Borchgrevink spent the first winter in the Antarctic. In the beginning of the last century a range of national expeditions were organized to explore the southernmost land of our planet. Among them were the Belgians, Germans, English, Swedish, French, Australians led by such worldly recognized polyarniks* as E. von Drigalski, R. Scott, E. Shackleton, J. Charcot, D. Mawson... All of them worked on the coasts of Antarctica without trying to penetrate into its interior. Than, in 1910–1913 two expeditions were undertaken to conquer the south geographic pole, one organized by a Norwegian R. Amundsen and the second one by an Englishman R. Scott. Unlike the one by Scott, the Amundsen's campaign was mainly a sports trip, it did not brought new information about the inland regions of Antarctica. On the contrary, during the English mission a comprehensive characteristic of the ice sheet surface was obtained including data on structure, thickness, mass-balance and movement of ice. A large contribution in improving the knowledge about Antarctic nature was made by American expeditions of the 30–40s organized by R. Byrd, as well as by French expedition to Terre Adelie in 1949–1951 and the Norwegian-British-Swedish one to Dronning Maud Land in 1949–1952.

* Russian word meaning "polar explorers"

In spite of all these efforts, only about 2 % of Antarctic surface was investigated by the middle of the XXth century thus making this continent a huge white spot of the Earth. This was one of the reasons of combining the logistic and scientific resources of many countries during the International Geophysical Year (1956–1958) for thorough and extensive exploration of the sixth continent. About 50 scientific stations were set up, many of them being located in the interior of East and West Antarctica. The main scientific results during IGY were obtained by American, Australian and Soviet Antarctic Expeditions. The Soviet one constructed three coastal bases: Mirny, Oasis and Lazarev, as well as five inland stations: Pionerskaya, Vostok-1, Komsomolskaya, Vostok and Sovetskaya. Starting from 1956 a number of inland traverses were undertaken. The most interesting results were brought by the traverses Mirny–Pionerskaya–Vostok, Mirny–Sovetskaya–Pole of Inaccessibility and Vostok–South Pole during the 2nd, 3rd and 4th Soviet Antarctic Expeditions, correspondingly.

On the 16th of December 1957 the convoy of tractors under the command of A.F. Treshnikov having left Mirny two months before and consisting of 9 vehicles reached the south geomagnetic pole in the point with the coordinates 78°28' S and 106°48' E where the new station was set up. The first wintering personnel of this station named "Vostok" (after the sloop of F.F. Bellingshausen) consisted of 9 persons, and the first chief of Vostok was V.G. Averianov. Complex meteorological and balloon-sounding observations were launched supplemented later by vertical ionosphere sounding, geomagnetic measurements, observations of aurora, measurements of ozone content and spectral properties of atmosphere, observations of cosmic rays and medical studies [Savatyugin, Preobrazhenskaya, 1999; Treshnikov, 1973].

In 1970 the glaciological-drilling group headed by N.I. Barkov initiated the regular snow-measuring observations and started the deep drilling of ice sheet. The coordinator of these works was Arctic and Antarctic Research Institute (AARI) in St. Petersburg (Leningrad at that time) and the drilling was carried out by the specialists of Leningrad Mining Institute. The drilling was stopped in February 1998 on the record depth of 3622.8 m. The study of the ice core samples has been performed by the joint efforts of Russian, French and American scientists. In the coming years further deepening of the borehole and study of the accretion ice samples is planned, followed by the penetration to Lake Vostok.

General characteristic of the area of study

The onset of the glaciation in Antarctica dates back to Miocene when the Antarctic circumpolar current formed, which limited the heat transport from the equatorial zone. The cooling was followed by the widening of snow cover which gradually turned into ice cover. Increased albedo favored further cooling of the Antarctic thus intensifying the meridional temperature gradient and strengthening the atmospheric circulation. Stronger moisture transport led to the growth of the new-born ice sheet. With time, the anticyclone circulation developed, and moisture flux from ocean to the continent stabilized. Unlike the continental glaciations of Northern Hemisphere, the Antarctic ice sheet after its formation has not disappeared during the climatic optima, only varying in size. The reason for this is the geographical location of the Antarctic responsible for the extreme conservatism of its climate and its resistance to the external influence [Averianov, 1990].

At present the surface of Antarctica is mainly covered by ice, with only few exceptions. The area of ice cover is 13.9 million km², 12.4 million of them is ground-based ice and about 1.5 million are ice shelves. The mean altitude of the continent's surface is about 2000 m, and the mean ice thickness is nearly the same. The maximum thickness of ice reaches almost 5 km. The total volume of ice imprisoned in Antarctica is about 24 million km³. If totally melted, this volume of ice would increase the level of the world ocean by about 70 m.

Based on the geological and geomorphological characteristics Antarctica is traditionally subdivided into two parts: East Antarctica (area 10.6 million km²) and West Antarctica (3.3 million km²). East Antarctic ice sheet is a vast, high and thick massif which comprises more than 4/5 of the total Antarctic ice volume. West Antarctic ice sheet has 5 times less area, it's 2 times lower and thinner. It is more pronouncedly subdivided into geomorphological forms of lower ranks. The differences in geographic location and relief of West and East Antarctica cause the climate of the latter to be much more severe [Averianov, 1990].

The present study is limited to the region of central Antarctica. According to Averianov (1990), the climatic region of Central Antarctica occupies the high plateau of East Antarctic ice sheet with the altitudes of higher than 2000–3000 m. The climatic conditions are characterized by the data from the stations located near its boundary (South Pole, Vostok-1) and in central parts (Komsomolskaya, Sovetskaya, Dome C, Vostok, Dome Fuji, Plateau). The most representative, because of its location and period of observations, is Russian Vostok Station

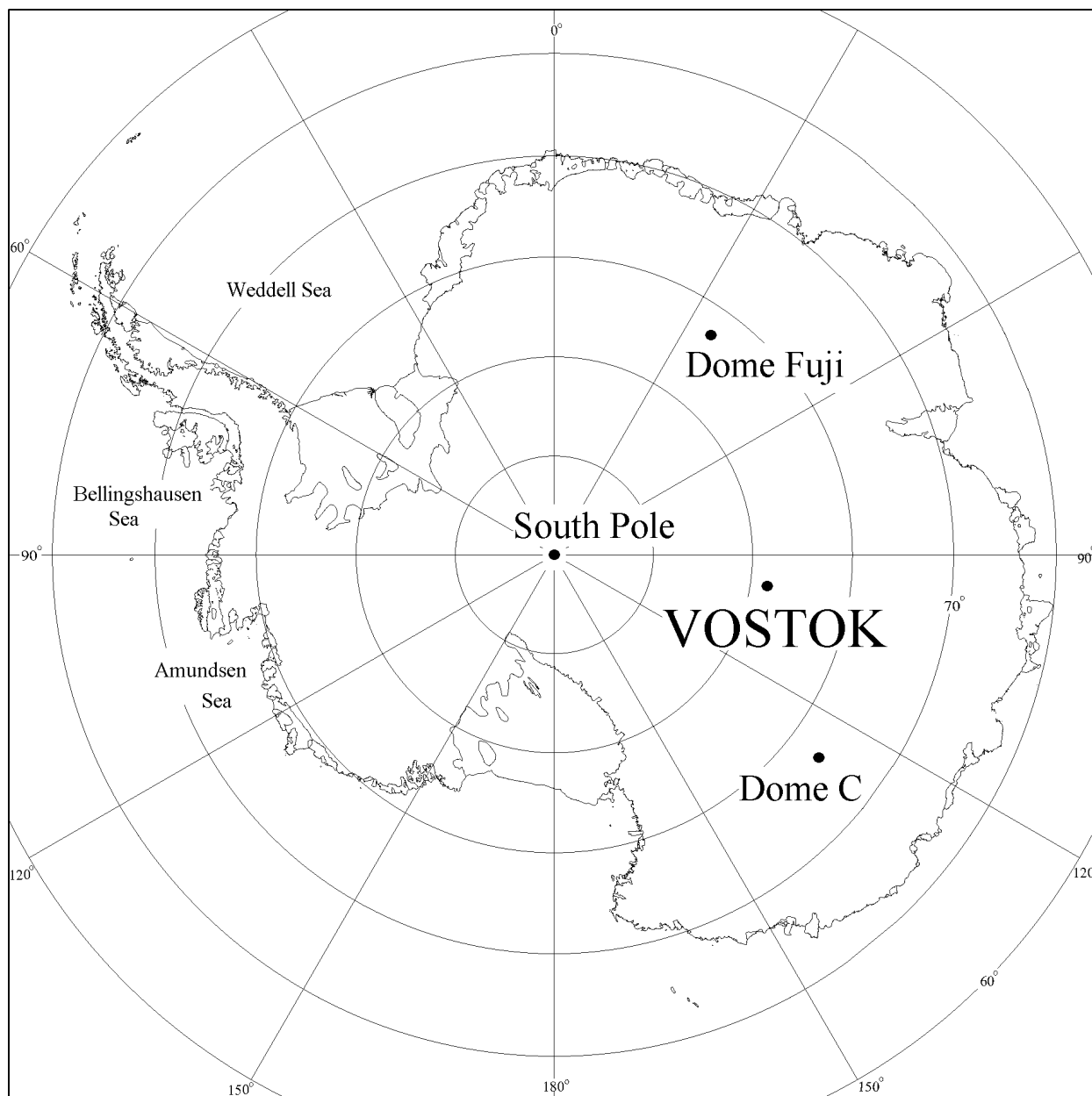


Fig. 1. Map of Antarctica with the location of Vostok Station.

situated in the point with the coordinates $78^{\circ}28' S$ and $106^{\circ}48' E$ at the altitude of 3488 m (Fig. 1). This is why the present study is mainly based on the data from this station.

Vostok Station (Fig. 2) is located in the central part of East Antarctica to the south-west from the main ice ridge dividing the basins of the Pacific and Indian oceans. The underlying surface in the vicinities of the station represents a plateau with a mean slope of less than 10^{-3} and characterized by the absence of large forms of ice relief. The glacier surface is covered by snow all year round which does not melt even during the warmest months.

The climate of the station is determined by its high-mountain and high-latitude location, as well as by remoteness from the oceans (the distance to the nearest coast is about 1300 km).

These factors make it difficult for the cyclones formed at the Antarctic and polar fronts to reach this area, this is why the most of the year the cold and dry Antarctic air mass is dominating here. The mean total cloud cover in central Antarctica is about 3.4, while lower clouds cover is close to 0. At Vostok, lower clouds are only observed few days per year. The most frequent are upper clouds (*Ci*, *Cs*) and middle clouds (*Ac*, *As*). Among the lower clouds, two forms can be observed: stratus (*St*) and strato-cumulus (*Sc*) [Spravochnik po klimatu, 1977].



Fig. 2. General view of Vostok Station (photo of Arctic and Antarctic Research Institute).

Because of this large number of sunny days, the incoming solar radiation is very high in central Antarctica and even exceeds that in the equatorial zone of Earth. At Vostok, annual sum of solar radiation is $4.6 \cdot 10^9 \text{ J m}^{-2}$, with 76% of this amount being direct radiation. Another reason for such an intense radiation is high transparency of the atmosphere and low water vapor content. However, because of high albedo of snow surface (0.82–0.86), most of the radiation is reflected back to the atmosphere, and the absorbed radiation is only $0.71\text{--}0.75 \cdot 10^9 \text{ J m}^{-2}$. The effective long-wave radiation of the surface is about $0.8 \cdot 10^9 \text{ J m}^{-2}$ and exceeds the incoming solar radiation. Thus, the annual radiation balance of the snow surface is negative and equals $-0.08 \cdot 10^9 \text{ J m}^{-2}$. In seasonal cycle, during the four months (from November to February) the balance is positive, while during the rest of the year the surface loses heat [Averianov, 1990; Rusin, 1961; Spravochnik po klimatu, 1976, 2002].

Because of radiation cooling, the temperature of the snow surface and near-surface air can reach very low values. According to the long-term observations, the mean annual surface air temperature at Vostok is -55.4°C , while mean temperature in winter is -66.2°C and in summer is -32.6°C (see the official web-site of AARI: www.aari.nw.ru). On the 21st of July 1983 the

coldest surface air temperature of our planet was registered here: -89.2°C . The mean annual temperature amplitude is about $30\text{--}40^{\circ}\text{C}$. The seasonal cycle of temperature is characterized by the phenomenon of "coreless winter" when there is no cold core of temperature during winter months due to warm air advections. The vertical profile of temperature is characterized by strong near-surface inversions that are observed practically all year round. The mean annual thickness of inversion layer is 730 m and the temperature difference between its upper and lower boundary is 18°C [Tsigel'nitsky, 1982]. Cold air is flowing down along the slope of the ice sheet thus forming "katabatic", or "inversion" winds. This is compensated by the inflow of relatively warm air in the free atmosphere from the coast into the interior of the continent [Voskresensky and Lysakov, 1976].

The mean annual air pressure at the station level is 625 mb (site AARI). The seasonal cycle of pressure is consistent with that of temperature: in winter the pressure is reduced due to air escaping to the lower atmosphere because of its seasonal cooling.

As mentioned above, the air in central Antarctica contains very little moisture: the mean annual water vapor pressure at Vostok is 0.07 mb, varying from ~ 0 mb in winter to 0.29 mb in summer. At the same time, due to the very cold temperature, the relative humidity (relatively to the saturating water vapor pressure over water surface) is comparatively high (70 %) and changes weakly from winter to summer [Spravochnik po klimatu, 1977]. Moreover, since the saturating water vapor pressure over ice is less than that over water, the atmosphere over Antarctic plateau is saturated (or even supersaturated) by moisture in relation to ice, which favors formation and growth of ice crystals. Taking into account very rare precipitation from clouds in central Antarctica, the mass of ice crystals falling from clear sky can represent a large proportion of the total annual amount of precipitation [Averianov, 1990]. However, the latter value is not known accurately because of obvious methodological difficulties in measuring such precipitation. The most likely value of the overall modern annual precipitation is between 25 and 40 mm.

The wind regime is characterized by katabatic winds which speed is proportional to the surface slope and inversion strength. Since the first factor is more important, mean annual wind speed is decreasing towards the interior of Antarctica reaching the value of 5.4 m s^{-1} at Vostok. In winter time the wind is stronger, which is related to more intensive inversion. Also, katabatic wind is known for the high stability of its direction that is related to the direction of the surface slope. In particular, at Vostok wind blows predominantly from WSW and WS direction [Spravochnik po klimatu, 1977]. Aside from this, cyclonic winds can be observed during the

invasion of warm air masses with speed and direction related to the spatial gradient of air pressure.

Since the glacier surface in central Antarctica is always covered with snow, blizzard is very common here. Obviously, the intensity of blizzard is wind speed dependant. For Vostok, typical wind speed during "near-surface" blizzard (according to Russian terminology[†]) is 6–7 m s⁻¹, and for "lower" blizzard is about 8 m s⁻¹. The frequency of all types of blizzard here is about 14 %, including 4 % for lower blizzard and upper blizzard (blizzard with snowfall) together [Averianov, 1972].

The mass balance of the snow surface is determined by two terms (taking into account the absence of melting and that snow blowing does not change long-term average snow accumulation): amount of precipitation and evaporation. Both of these values are not known precisely for Vostok [Averianov, 1990], but the uncertainty of the mass balance value itself is much less because it can be determined directly by means of snow-measuring observations. In particular, according to the stake measurements on the stake network, the mean annual snow accumulation rate in the vicinity of Vostok equals 2.2 g cm⁻² year⁻¹ [Barkov, Lipenkov, 1996].

The properties of snow cover are formed by such factors as solar radiation, temperature and wind regime as well as snow accumulation rate, which determines the intensity of snow metamorphism processes. The region of central Antarctica entirely belongs to the snow zone of ice formation for which very gradual changes of firm properties with depth are typical [Averianov, 1990; Kotlyakov, 1961, Petrov, 1975]. In the vicinity of Vostok the seasonal differences of snow cover are absent mainly due to extremely low annual snow accumulation (about 7 cm in snow equivalent, which is about 2 times less than average amplitude of micro-relief). As a rule, a stratigraphic layer representing one year of accumulation is marked on the top by a radiation crust with an underlying thin depth hoar layer. There are often dense wind layers formed as a result of snow re-deposition by wind. They can sometimes substitute several annual layers. The average density of the surface (20 cm) snow layer at Vostok is 0.32 g cm⁻³ [Barkov, Lipenkov, 1996].

The data on the main glacio-climatic characteristics of central Antarctica are summarized in Atlas of Antarctic [1966, 1969] and Atlas of snow and ice world resources [1997].

[†] "Near-surface blizzard" ("pozemok" in Russian) is characterized by snow blowing in the near-surface (few cm) air layer; "lower blizzard" is observed during stronger wind and snow particles are lifted to greater height but usually less than the level of observer's eyes; "upper blizzard" is observed when blizzard is accompanied by snowfall.

The aims of the study

As pointed out above, one of the priority scientific activities in Antarctica is the study of physical and chemical properties of ice cores obtained by deep ice drilling. One of the main parameters of interest is the isotopic composition of ice deposits which is a paleo-temperature indicator (see section I.2). To calibrate this isotope thermometer, detailed study of present-day spatial and temporal variability of snow isotope composition is necessary. Thus, the main aim of the present work is to determine the relationship between local air temperature and snow isotope composition at Vostok Station, and to reconstruct climatic variability in this area over the past 200 years. To achieve this goal, the following tasks had to be solved:

1. To carry out the field study of spatial and temporal variability of snow isotope composition and accumulation rate in the vicinity of Vostok Station;
2. To determine the main factors responsible for the formation of snow isotope composition in central Antarctica;
3. To study the meteorological conditions of precipitation formation at Vostok;
4. To study the role of snow redistribution processes in spatial and temporal variability of its isotope composition;
5. To reconstruct changes in air temperature and snow accumulation rate at Vostok over the past 200 years using the newly obtained results and stratigraphic and geochemical data from deep pits.

I. METEOROLOGICAL REGIME AND ISOTOPE COMPOSITION OF PRECIPITATION: REVIEW OF LITERATURE

I.1. Meteorological regime and precipitation formation in central Antarctica

Meteorological regime of central Antarctica forms as a result of interaction of a number of factors among which the most significant are radiation balance of the underlying surface (briefly discussed in the Introduction) and atmospheric circulation that brings heat and moisture to the central part of the continent.

I.1.1. Atmospheric circulation in the southern hemisphere.

The latitudinal temperature gradient existing all year round in the whole troposphere above Antarctica is responsible for the formation of the circumpolar vortex in the free atmosphere which is characterized by lower pressure in its center and clockwise rotation. As a result, air is descending over the most of the continent [Voskresensky, Lysakov, 1976], which is one of the main factors in forming the anticyclonic type of weather. The predominance of clear sky typical for such weather is favorable for the radiative cooling of the surface, while low temperature causes an extreme dryness of the air which leads to further cooling. Cold air is flowing down along the glacier slope, while in the free atmosphere this flux is counterbalanced by inflow of moist and warm air from ocean. The change of the direction of the meridional component of the air flux takes place at the altitude of 3.8–5 km above sea level. This circulation develops most intensively in winter when the gradient between the pole and the low latitudes is the strongest [Averianov, 1990; Schwerdtfeger, 1987]).

The above picture is often disturbed by meridional invasions into the high-latitude region of the cyclones formed at polar or, less often, Antarctic fronts [Dydina et al., 1976; Savitsky, 1976]. The latter are usually smaller and less developed in height. They are formed at the latitudes of 60–65 °S and the zonal component is dominant in their movement: they move around Antarctica parallel to the main stream, i.e., from west to east. Polar cyclones are generally deeper and larger than Antarctic ones. Possessing considerable meridional component in their movement, they sometimes penetrate far into the Antarctic plateau and thus play an important role in the inter-latitudinal exchange of heat and moisture of the southern hemisphere. Approaching the boundary between comparatively warm waters of the Southern ocean and cold Antarctic coast the polar cyclones can become stronger and, provided the presence of the blocking ridges of high pressure, stationary. On the maps of long-term average cyclone system

density the areas of most frequent cyclone occurrence can be clearly seen (Fig. 3): Weddell, Ross, Bellingshausen and Commonwealth seas. These areas play important role in formation of climatic regime of Antarctica [Averianov, 1990; Schverdtfeger, 1987].

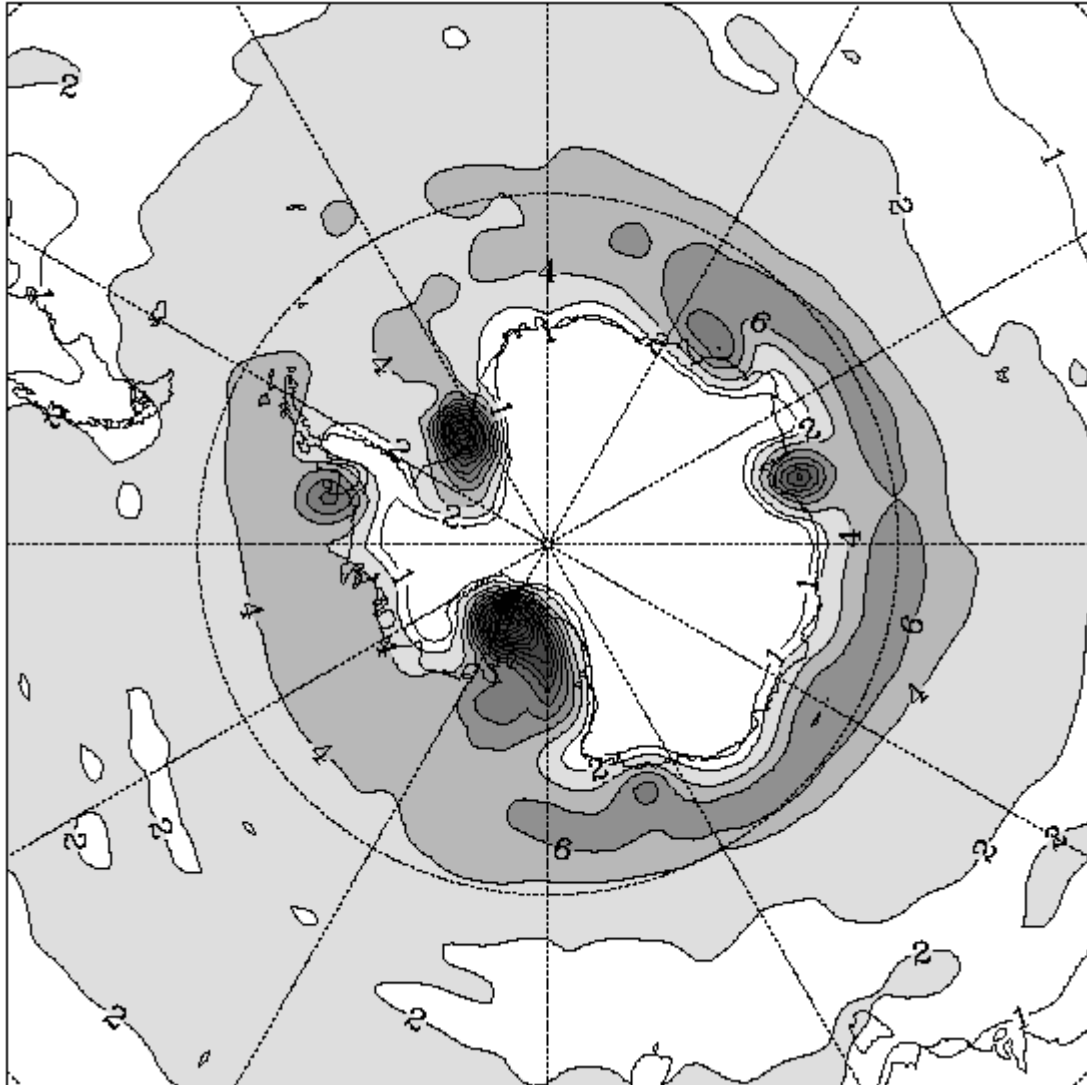


Fig. 3. Climatological cyclone system density distribution derived from the NCEP (National Center for Environmental Prediction) reanalysis (1958–1997) for winter (from [Simmonds, 2003]). The contour values are $1, 2, 4, 6$ and $8 \cdot 10^{-3}$ (degrees of latitude)².

As a whole, the atmospheric circulation in the middle and high latitudes of the southern hemisphere is governed by the following basic regimes (see the review in [Simmonds, 2003]).

First of all, this is the so-called Southern Annular Mode [Thompson, Wallace, 2000] that is characterized by opposite air pressure variations in the middle and high latitudes of the southern hemisphere. The index of its intensity is Antarctic Oscillation Index (AOI) representing the difference of mean latitudinal near-surface air pressure at 40 and 65° S [Gong, Wang, 1999].

Higher index means stronger gradient of pressure and temperature between high and middle latitudes, stronger westerly and weaker inter-latitudinal exchange, which causes cooling in Antarctic. This annular mode is related to the tropical circulation (ENSO), which is confirmed by the fact that El-Nino years are often correspond to the lower values of AOI [Maslennikov, 2002a,b].

Another important regime is the Antarctic Circumpolar Wave, which characterizes the drift of anomalies of meteorological and oceanographical parameters around Antarctica from west to east with a period of about 8–10 years [White, Peterson, 1996]. Anomalies of temperature and pressure, being born in the subtropical zone of the Pacific in relation to El-Nino, are then transferred by Antarctic circumpolar current to the east. This phenomenon is specific to the southern hemisphere, because in the northern one there is no continuous circumpolar current [Peterson, White, 1998]. The period of oscillations related to this wave is 4–5 years.

Disturbance to the two previous regimes is brought by Antarctic Dipole Mode (ADM) that is opposite oscillations of temperature, pressure and sea ice cover in east part of Pacific sector and in Atlantic sector of Antarctic [Yuan, Martinson, 2001]. The Antarctic Dipole is related to tropical circulation, too, the El-Nino years being characterized by positive anomalies of temperature in the Pacific sector and negative ones in Atlantic sector. ADM is actually one of the strongest mechanisms responsible for the transmission of the climatic signal from low to high latitudes [Liu et al., 2002].

Despite relatively weak degree of investigation of the above circulation regime, their role in forming climatic variability of interior parts of Antarctica is in general beyond doubt. In years of anomalous development of meridional processes more cyclones invade into the continent, which causes warming and increasing of precipitation. On the contrary, when zonal processes are stronger, air temperature and pressure are lower in high latitudes [Dydina et al., 1976; Zhukova, 1986; Savitsky, 1976]. In particular, in years with higher AOI index increased temperature is observed over Antarctic Peninsula and decreased over the rest of the continent, especially in East Antarctica. The influence of the tropical circulation on the Antarctic climate mainly reveals itself in reduced sea ice cover in Amundsen and Bellingshausen seas in the El-Nino years (which corresponds to negative Southern Oscillation Index) and to lesser degree in cooling of the interior part of the Antarctic [Kwok, Comiso, 2002]. Thus, cooling observed during the last 10–20 years over the most of the continent with simultaneous warming over the Peninsula [Doran et al., 2002] is consistent with stronger Southern Annular Mode and El-Nino during the same period. Rapid warming in the area of Antarctic Peninsula is related to stronger westerly and thus to more intensive advection of warm oceanic air, as well as with destruction of sea ice in the

surrounding seas [Kwok, Comiso, 2002]. Intensification of the annular mode (increasing of AOI index) is accompanied by an increased air pressure to the north of 40° S and its decrease in the high latitudes. At the same time smaller amount of cyclones are observed in the southern hemisphere. This apparent contradiction is explained by the fact that though the number of cyclones is less, they became deeper and more intense [Simmonds, Keay, 2000].

I.1.2. Surface temperature inversion at Vostok and wind regime.

Since the mean values of main meteorological parameters at the near-surface level were discussed in the Introduction, below we will consider the meteorological regime of troposphere using the published data of balloon-sounding observations.

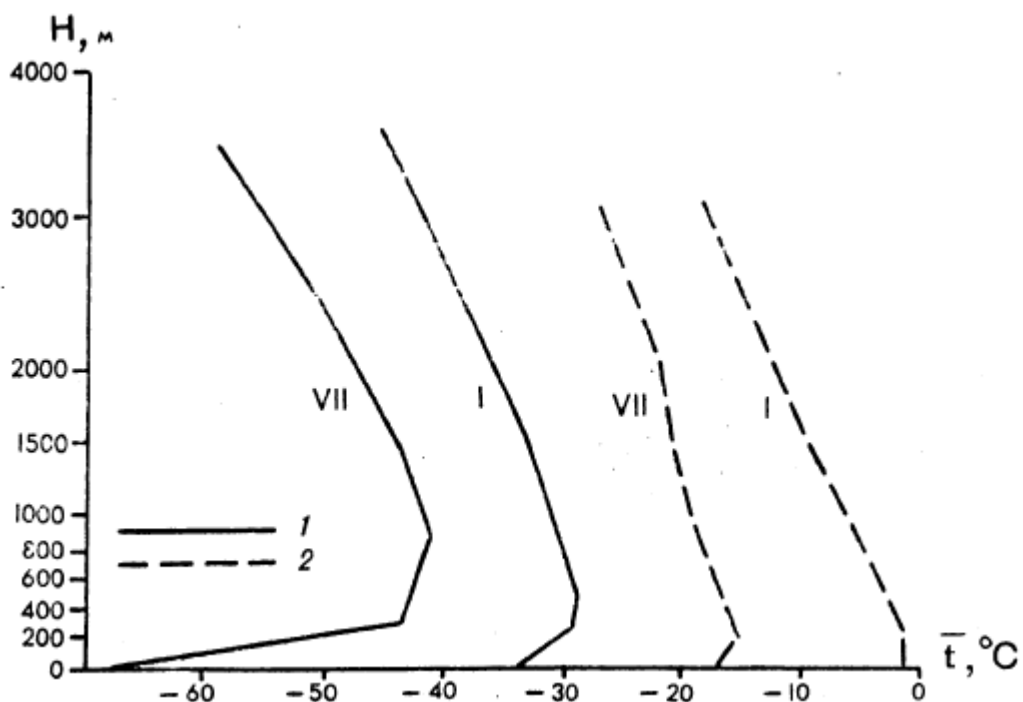


Fig. 4. Vertical distribution of air temperature in the boundary layer at Vostok (1) and Mirny (2) stations in July and January (from [Voskresensky, Tsigel'nitsky, 1985]).

The most typical feature of tropospheric structure in central Antarctica is a stable thick layer of surface inversion of mixed radiation and dynamic origin [Connolley, 1996; Phillpot, Zillman, 1970; Tsigel'nitsky, 1982; Voskresensky, Tsigel'nitsky, 1985]. The mean thickness of inversion in winter is about 800 m with a temperature difference between upper and lower boundary of about 25 °C and occurrence of nearly 100 % (Fig. 4). These values are twice of those in central Greenland. Monthly means of the main inversion characteristics are listed in Table 1.

Table 1. Mean values of thickness (h), temperature difference (ΔT) and occurrence (p) of surface inversion at Vostok for the period 1958–1961, 1963–1969 (from [Tsigel'nitsky, 1982])

Parameter	I	II	III	IV	V	VI	
h (m)	540	790	820	790	790	780	
ΔT (°C)	4.3	12.6	21.0	24.7	24.4	24.2	
p (%)	94	100	100	100	100	100	
Parameter	VII	VIII	IX	X	XI	XII	Year
h (m)	770	770	780	760	710	420	730
ΔT (°C)	25.0	25.5	24.4	18.2	8.8	3.2	18.0
p (%)	100	99	99	99	90	95	98

During its maximum development, in winter, the inversion layer is not homogeneous. Three sub-layers can be distinguished. The first one is about 100 m thick and characterized by the most intensive temperature changes with mean vertical gradient of $-8\text{ }^{\circ}\text{C } 100\text{ m}^{-1}$. For the second, that has thickness of 250 m, a weaker gradient ($-2\text{ }^{\circ}\text{C } 100\text{ m}^{-1}$) is typical. Finally, the third one (about 500 m) is isothermal. Just above the surface inversion a quasi-stationary layer is situated with weak positive gradients of temperature ($0.3\text{ }^{\circ}\text{C } 100\text{ m}^{-1}$). Thus, during the periods of maximum inversion development the normal temperature distribution typical for free atmosphere ($0.6\text{ }^{\circ}\text{C } 100\text{ m}^{-1}$) is established only from the altitude of 2000 m above ice surface [Tsigel'nitsky, 1982; Voskresensky, Tsigel'nitsky, 1985].

With such high values of thickness, intensity and probability surface inversion plays a role of screen preventing thermal and dynamic interaction of free atmosphere with the underlying surface. It is confirmed by the fact that the maximum amplitude of temperature in winter in the isothermal layer above inversion ($11.0\text{ }^{\circ}\text{C}$) is twice as less as near the surface ($21.2\text{ }^{\circ}\text{C}$) and less than in free atmosphere ($13.1\text{ }^{\circ}\text{C}$) [Tsigel'nitsky, 1982].

In summer months because of radiation heating the thermal stability of the lower atmospheric layers above Antarctica sharply decreases. In the afternoon the surface inversion can be completely destroyed. Moreover, at this time of the day the conditions are favorable for the formation of a thin (about 100 m) layer with unstable stratification overlaid by elevated inversion or by isothermal layer [Tsigel'nitsky, 1967].

The inter-annual variability of inversion parameters (thickness and intensity) is a complex index of climate variability of central Antarctica because inversion is formed under the influence of several main climatic factors: underlying surface, radiation conditions and atmospheric circulation. During the period 1958–1982 the inversion parameters revealed significant trends

which were opposite for thickness and intensity of inversion. It is explained by the fact that, with an increasing (decreasing) inversion thickness, the gradient of temperature in the inversion layer will become weaker (stronger) provided that the intensity of radiation cooling remains the same. In general, during 1958–1982 the mean thickness of winter (June–August) inversion lowered by 330 m and ΔT increased by $1.02\text{ }^{\circ}\text{C}^*$. At the same time the surface air temperature warmed by about $1.25\text{ }^{\circ}\text{C}$. Since dynamic settling of air masses is important for inversion formation, reduced inversion thickness can be related with an increased intensity of vertical air movement. The growth of ΔT means a more rapid warming in the free atmosphere comparing to the near-surface air, which is explained by the screening effect of the inversion layer [Tsigel'nitsky, 1990].

The seasonal variation of temperature in troposphere, like near the ground surface, has coreless character. At all the altitudes the warmest month is January, the coldest are August and September. The temperature distribution in whole boundary layer has positive asymmetry, which is especially pronounced in winter. The reason for this is warm advection during strengthening of meridional circulation and cyclone invasions into the continent [Tsigel'nitsky, 1982].

The vertical distribution of wind is characterized by a rapid growth of wind speed in the lower inversion layer due to diminution of friction. Maximum speed is reached at the lower boundary of the isothermal layer and wind here is actually stronger than geostrophic wind. This phenomenon is called "meso-jet stream" [Vorontsov, 1967]. The origin of this wind is due to katabatic forces. During cyclonic weather situations, the speed and direction of the wind is highly variable and depends on the station's position relative to the center of cyclone. Seasonal variations of wind speed in the boundary layer are characterized by higher values in winter due to both stronger inversion and more frequent cyclone invasions [Averianov, 1990; Tsigel'nitsky, 1982; Voskresensky, Tsigel'nitsky, 1985].

The height of boundary layer in central Antarctica is determined as 1) the mean altitude at which wind rotation stops (for dynamic boundary layer) and 2) the upper boundary of surface inversion in winter or elevated inversion in summer (thermal boundary layer). At Vostok the height of the dynamic boundary layer is 660–670 m in winter and 1250–1400 m in summer and that of thermal one is 610–650 m in winter and 1500–1800 m in summer (see Table 5 from [Tsigel'nitsky, 1982]).

I.1.3. Precipitation and water vapor in central Antarctica.

The influence of the main factors governing formation and precipitation of atmospheric moisture (humidity and temperature of air masses, atmospheric circulation) is controlled in

* Absolute values of these changes are likely too high because of inhomogeneity in the series of balloon-sounding data (see Chapter "Experimental data", section II.1).

Antarctica by the elevation of the cold ice sheet surface, the distance from the ocean, and the location in relation to the major cyclonic paths. Different combinations of precipitation-forming factors with morphometric features of the ice sheet preventing moist oceanic air penetration into the interior of the continent are favorable for formation of three types of precipitation. They are: snow from clouds, ice crystals from clear sky and rime on the surface (see the review in [Averianov, 1990; Petrov, 1975; Schwerdtfeger, 1987]).

According to the ideas formed in the first half of the last century [Shumsky, 1955] the conditions of formation, growth and precipitation of ice crystals in the atmosphere, as well as their forms and sizes, are related to temperature and humidity of air. The necessary condition of crystal formation and growth is supersaturation of air by water vapor which can be reached mainly by cooling. The nuclei of condensation are usually marine aerosols brought by marine air masses [Hogan, 1997; Golubev, 2000].

The two main types of atmospheric crystals are lamellar (growing to the direction of basic plane) and columnar (growing to the direction of main crystal axis). Both of them have a great variety of sub-types. During the snowfalls different sub-types (and types) are usually aggregated. Columnar crystals are often smaller than lamellar ones. Besides, there are two other types of crystals: needle-shaped and grains. It is assumed that needles are growing in the direction of a secondary axis and thus can be considered as asymmetric lamellar crystals. Snow grains, sleet, are formed as a result of supercooled water freezing on the surface of ice crystals [Averianov, 1990].

High supersaturation of cold air leads to the formation of crystals with complex shapes, like stars or dendrites. Low supersaturation leads to more regular shapes. The colder is the air, the higher is the ratio of the columnar crystals, and the smaller are the crystals of all the types [Averianov, 1990; Bromwich, 1988; Golubev, 2000].

In central Antarctica, up to 98 % of the total precipitation is formed by columnar crystals with a typical length of 0.025–0.6 mm and thickness of 0.01–0.08 mm. Most of the snow falls from *As* and *Ac* from the height of 1000–3000 m above ice sheet surface [Averianov, 1990].

The mean total water content of the atmosphere in the vicinity of Vostok (for clear sky conditions) for the period 1977–1981 is 0.34 mm. Seasonal changes of this parameter is comparatively simple and closely related to those of air temperature: from 0.17–0.19 in winter to 0.73–0.74 in summer (Fig. 5) [Burova et al., 1990]. The seasonal variability of atmospheric water content is confirmed by satellite observations on total water vapor content [Miao et al., 2001]. At the same time, the inter-annual variability of this parameter is practically absent. In particular, during the period 1960–1985 the mean total summer water content at Vostok did not

changed while at most coastal stations it increased, which is related to warming observed during this period [Burova, Voskresensky, 1990]. At the same period of time, the precipitation rate has increased by about 5 %, which is explained by intensification of cyclonic activity [Bromwich, Robasky, 1993].

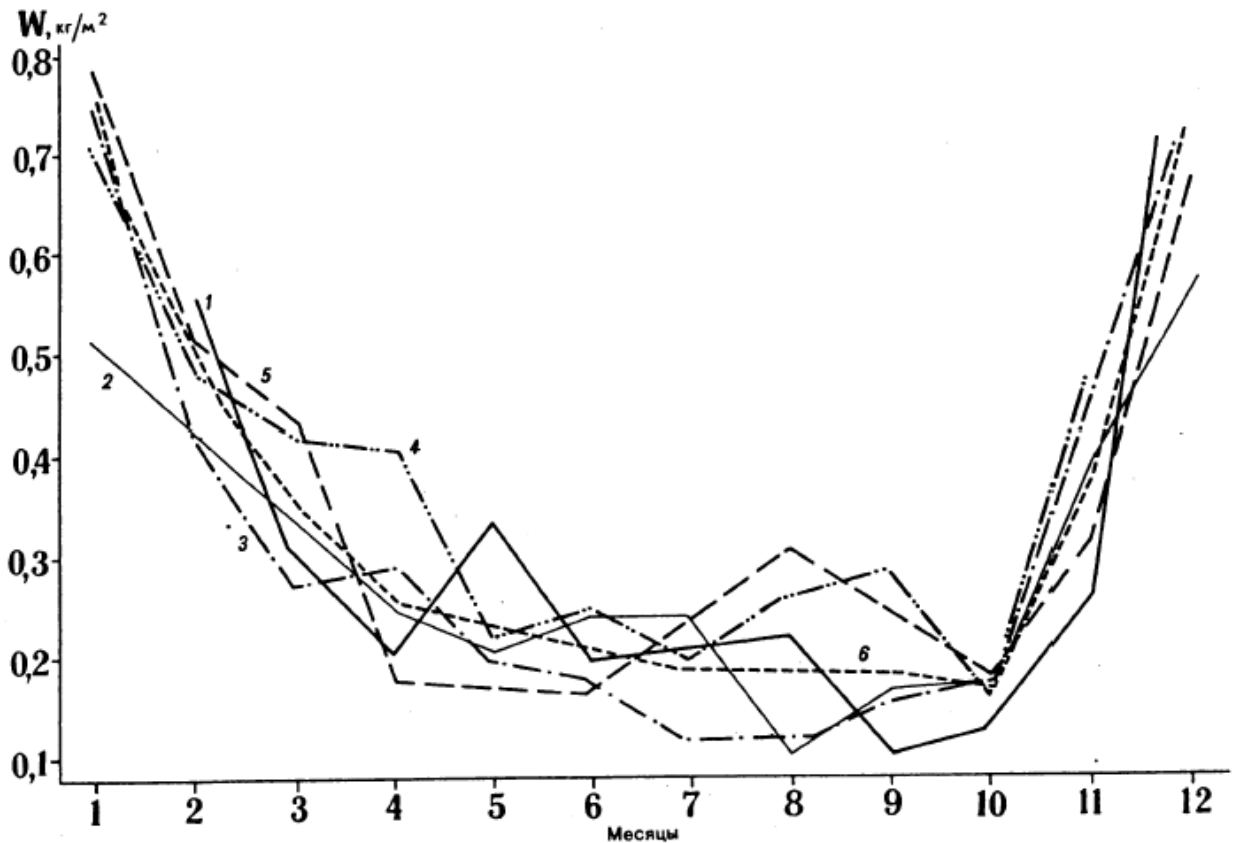


Fig. 5. Seasonal variation of total water content of atmosphere above Vostok (in kg m^{-2}) during the days with clear sky (0–3 balls): 1–5 – correspondingly, 1977–1981, 6 – mean value for 5 years (from [Burova et al., 1990]).

Snow from clouds falls during passage of cyclones over the ice cover, so it is frequent at the coast and over the lower part of the ice sheet slope. Cyclonic precipitation at the coast forms in alto-stratus, strato-nimbus and less frequently in stratus clouds. The most frequent and intense snowfalls occur over the areas where local altitude does not exceed the level of condensation, i.e., 800–2000 m above sea level [Averianov, 1990; Aleksandrov et al., 1991; Kotlyakov, 1961]. About 20 % of Antarctica lies below this height but this part of the continent receives roughly one half of all the snow deposited on the whole ice sheet [Averianov, 1990]. To the central parts, the frontal clouds carrying precipitation penetrate quite rarely. And even if it happens, the clouds are already depleted in moisture, so the snowfalls are less intensive than over the coast. The

number of days with snow from clouds per year in Antarctica is as follows: about 150 on the coast, 200 over the lower part of the slope and 25–50 in the central part [Bryazgin et al., 1990; Spravochnik po klimatu, 1977].

The question about the source of moisture feeding the area of Vostok is still under discussion. Results of General Circulation Models simulations suggest that most of the precipitation (40 %) at Vostok comes from low and middle latitudes of the Indian ocean, about 10–20 % does from the Atlantic and the Pacific, as well as from Antarctic seas, and the rest (less than 10 %) comes from the Antarctic ice shelves. The contribution of local moisture is negligibly small [Delaygue et al., 2000]. On the other hand, experimental data lead to another conclusion. The results of chemical analyses of snow cover imply that Vostok source area is in the Pacific sphere of influence [Averianov, 1969]. The same conclusion is achieved when analyzing the distribution of isotope composition of surface snow in Antarctica: on the diagram of snow isotope composition versus mean site annual temperature Vostok Station is situated on the continuation of Patriot Hills–South Pole line (Pacific sector) and aside from lines Mirny–Komsomolskaya and Dumont-d'Hurville–Dome C (Indian sector) [Ekaykin et al., 2001] (see also Fig. 8 in this work).

The precipitation of tiny ice crystals under clear sky conditions is typical for the anticyclonic weather which is dominant in the interior parts of the continent. This phenomenon is sometimes called "ice crystals" or "diamond dust". Deposition of these ice crystals is constantly observed at the high-latitudinal Antarctic stations: 247 days per year at Vostok [Averianov, 1972] and 316 days per year at Plateau [Schwerdtfeger, 1987].

The formation of ice crystals from cloud-free sky is related to the seeping of moist air into the cold continent at altitudes of about 500–1000 m above its surface followed by supersaturation of air by water vapor relative to ice due to radiation cooling [Voskresensky, 1976; Voskresensky, Lysakov, 1976]. Cooling proceeds with a rate of $2\text{ }^{\circ}\text{C day}^{-1}$ over the inversion and more intensively, up to $4\text{ }^{\circ}\text{C day}^{-1}$, in the inversion layer itself. Over the central parts of the ice sheet the ice crystals are born in the relatively warm isothermal layer just above the surface inversion. At South Pole this layer is situated between the levels of 650 and 600 hPa, at Vostok between 650 and 550 hPa [Artemiev, 1976; Averianov, 1990; Bromwich, 1988; Robin, 1977; Schwerdtfeger, 1987]. The formation of ice crystals in the inversion layer is confirmed by the fact that considerable amounts of water vapor are contained here (on average, 30 % of the total amount of vapor in the column from ground to the height of 8 km) [Burova et al., 1990].

At Plateau Station, during a year there are 51 days with snow from clouds that give 13 % of total annual precipitation, while the remaining 87 % are brought by ice crystals from clear

sky. In 1967 and 1968, during 5 winter months, when no snowfalls were observed, the accumulation of ice crystals occurred with an average rate of $1.2 \text{ cm month}^{-1}$. At the same time, according to theoretical estimations, precipitation from clear sky is equal to about $1 \text{ g cm}^2 \text{ year}^{-1}$ [Radok, Lile, 1977]. Generally, theoretical considerations give very different values for the annual amount of ice crystals: from 0.28 to $2.5\text{--}3 \text{ g cm}^{-2}$ [Averianov, 1990]. Thus, contribution of this precipitation type could be significant, but its value is still not known with satisfying accuracy.

Concerning precipitation forming directly on the surface of snow (rime), both theoretical estimations and experimental measurements are available. According to the estimations by turbulent diffusion, condensation (inverse sublimation) exceeds sublimation from March to October giving the total sum of rime equal to 0.018 g cm^{-2} [Artemiev, 1976]. Direct measurements were made in 1982 by D.N. Dmitriev who weighted a cup (evaporator) with a sample (frozen water) once a day to know how much material had sublimated or condensed. He found that inverse sublimation dominated from March to September and its total sum was 0.06 g cm^{-2} , or 3 % of annual accumulation. This value is roughly 3 times higher than the above mentioned theoretical one. Similar measurements were performed earlier (in November 1961 – January 1962) by V.K. Nozdryukhin. As evaporator he used a plastic cup filled with a monolith sample of surface snow, and the weighting was carried out twice a day, at 7 a.m. and 7 p.m. of local time. The results suggested that during the night half of polar day condensation exceeds sublimation. On average, night-time condensation amounts to 15 % (in November), 31 % (in December) and 33 % (in January) of day-time sublimation [Averianov, 1990]. On the contrary, measurements fulfilled at Japanese Dome Fuji Station showed that during period from March to October monthly sums of condensation was about an order of magnitude larger than at Vostok [Motoyama, personal communication, 2003].

Weather conditions accompanying different types of precipitation do not generally correspond to the mean annual ones. During cloud snowfall, surface air temperature is some $5 \text{ }^\circ\text{C}$ higher, wind is stronger, air humidity is elevated, total cloud cover is 2 times higher, probability of middle and upper clouds is increased by 50 % and that of clear sky decreased by about 80 % compared to the annual mean values. Precipitation of ice crystals is characterized by air temperature and wind speed lower than average, while the other parameters are roughly equal to their annual means.

Rime is observed during the weather conditions close to those of ice crystals, which is explained by the fact that these two types of precipitation often take place simultaneously [Averianov, 1972].

It has been repeatedly noted [Averianov, 1990; Bromwich, 1988; Bryazgin, 1990; Kotlyakov, 1961; Rusin, 1961; Schwerdtfeger, 1987] that in polar regions the results of direct measurements of solid precipitation are disturbed because of imperfection of instruments and methods. This is first of all related to wind activity which is able to blow snow from and into precipitation collector. Due to this reason many national Antarctic expedition do not perform such measurements at all. If the precipitation data are used for climatological studies, they can sometimes lead to false conclusions. To correct data on precipitation rate, correction coefficients were suggested that account for the influence of wind [Bryazgin, Voskresensky, 1991] and a technique was developed to estimate the amount of precipitation from the data of its duration and intensity [Bryazgin, 1990]. Since recently, precipitation modeling has being developed actively with use of meso-scale and general atmospheric models [Bromwich et al., 1995; Cullather et al., 1998; Gallee et al., in preparation, Genthon et al., 1998].

Precipitation amounts can also be determined by water balance method which gives an error of $\pm 20\%$ for Antarctica. Such an uncertainty could be sufficient for glaciological purposes but only few sites have been monitored for the data necessary for the calculations. In particular, in central Antarctica annual precipitation equals to the sum of surface mass balance and sublimation, taking into account that melting is zero and mass balance due to snow drift is on average nil, too [Averianov, 1990; Petrov, 1975]. While mass balance (snow accumulation rate) can be determined sufficiently accurately [Barkov, Lipenkov, 1996; Petrov, 1975], the value of sublimation is known with a large uncertainty. For Vostok, calculations based on turbulent diffusion [Artemiev, 1976] and direct measurements [Averianov, 1990] give similar values and suggest that the sum of sublimation during the warm period equals to 2–4.5 mm. Thus, summer sublimation removes about 10–20 % of annual precipitation amount, and the latter is thus equal to 24–27 mm year⁻¹. On the other hand, direct measurements at Plateau Station showed that summer snow sublimation value is about 18 mm, that is, 4–9 times larger than at Vostok [Radok, Lile, 1977]. At the same time, at Dome Fuji Station the sublimation rate was measured to be roughly two times lower than at Vostok [Motoyama, personal communication, 2003].

It should be however noted that these estimations do not take into account sublimation of drifting snow which can be an important term of surface mass balance and heat budget [Bintanja, Reijmer, 2001; Gallee et al., in preparation].

I.2. Isotope composition of precipitation and its relation to the conditions of formation: Theoretical considerations and empirical data

The term "isotope" was firstly suggested by English physicist F. Soddy in 1910, although the idea about elements that have the same charge but different atomic mass had appeared long before. Already 8–9 years after this event the first measurements of concentration of stable isotopes were carried out. In 1929 and 1932 heavy isotopes of oxygen and hydrogen were discovered by Giauque and Johnston (1929) in Great Britain and Urey with others (1932) in the United States. During the following 35 years an intensive work had been undertaken to study the processes governing the natural distribution of isotope composition of these two elements. The results of these efforts were summarized in a paper "Stable isotopes in precipitation" of a Danish scientist W. Dansgaard that came out in 1964 [Dansgaard, 1964].

The first idea of using isotopes as a natural paleothermometer belongs to Urey who found a small difference in the isotopic composition of carbonates depending on the temperature of calcite formation. The first relationship between isotope composition of water precipitation and temperature of condensation was described by Dansgaard. These discoveries set up a basis for the method of paleotemperature reconstructions by isotope analyses of fossil precipitation samples. It was soon realized that one of the best application for the water stable isotope geochemistry was the study of the isotope composition of past snow precipitation successively accumulated during many millennia in polar ice sheets. The ice cores obtained by deep drilling of glaciers represent unique archives of climatic information in which the data on past changes of temperature, snow accumulation rate, wind speed, chemical and gas composition of atmosphere are stored. The first deep drilling of polar ice was completed in 1964 at Camp Century site in Greenland [Dansgaard et al., 1971]. The 1390-m deep borehole reached bedrock, and the analysis of about 1600 ice samples allowed reconstructing climate in this area over 100 ka. In Antarctica the first deep drilling project was fulfilled in 1966 at American Byrd Station [Epstein et al., 1970]. In 1974 and 1978 the first boreholes were finished at Vostok Station [Barkov, 1970; Barkov et al., 1975] and French Dome C Station [Lorius et al., 1979]. At present, deep drilling projects are carried out in various sites in Arctic and Antarctica by specialists from Europe, the United States, Russia, Japan and other countries. In February 2003 the hole at Dome Concordia site drilled in the frame of EPICA (European Project of Ice Coring in Antarctica) reached the depth of 3200 m. The age of the deepest ice obtained from this borehole can be as old as 800 ka (personal communication of Jean Robert Petit, 2003). In July 2003 the head of the drill reached bedrock at NorthGRIP site (central Greenland) at the depth of 3085 m thus making this borehole

the longest one ever drilled on this island. The studies of the obtained ice cores including the measurements of their isotope composition have substantially broadened our knowledge on the climatic changes in polar regions and the whole Earth during the last half million of years.

I.2.1. Theoretical basis of the relationship between isotope composition of precipitation and air temperature: Simple isotope models and GCMs

The relationship between isotope composition of precipitation and temperature of its formation is based on the "isotope depletion" of moisture in the precipitating air mass due to isotope fractionation at each phase change. Since saturation water vapor pressure is less for heavy water molecules (HD^{16}O and H_2^{18}O) than for light molecules (H_2^{16}O), the concentration of heavy isotopes in the liquid phase is higher than in the vapor phase equilibrated with this liquid. So, the isotope composition of water vapor contained in an air mass formed over the ocean is negative (if expressed in δ notation[†] – see equation (2)). As the cooling of the air mass proceeds, the water vapor condenses and new portions of precipitation are enriched in heavy isotopes in relation to the vapor remaining in the air mass thus making the vapor more and more isotopically depleted (Fig. 6). Obviously, in the course of further cooling both vapor and condensate become isotopically lighter due to the washing out of heavy water molecules during precipitation formation.

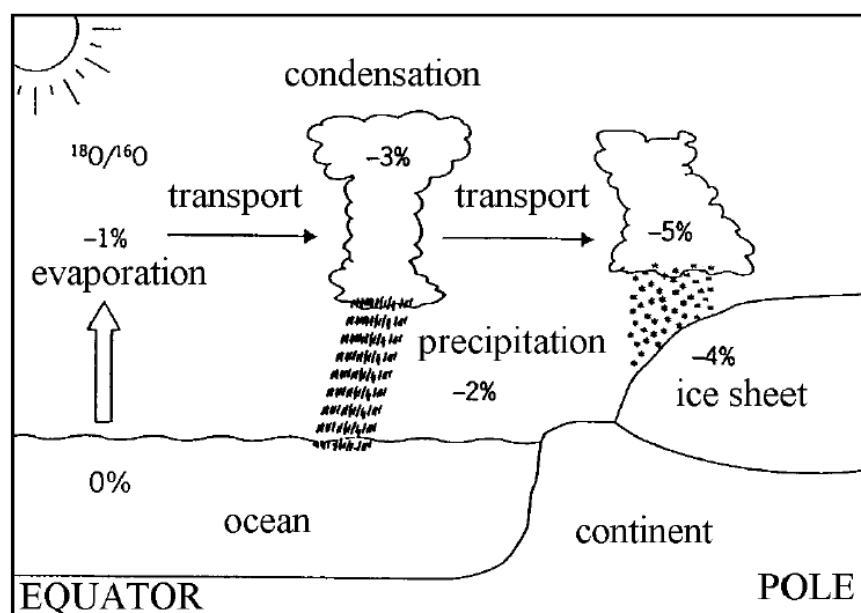


Fig. 6. Natural water cycle and isotope fractionation (from [Joussaume, 1993]).

[†] See the list of basic notations and abbreviations in Annex A

At any moment of time the isotope composition of precipitation is thus primarily determined by the ratio F of moisture contained in the air mass to its initial moisture content. In turn, F depends on the difference of condensation temperature between the current time and at the beginning of the distillation process. Isotope composition of liquid precipitation (δ_p) can be fairly well expressed by a Rayleigh-type equation which is based on the assumption that condensation takes place in the dynamic and isotopic equilibrium and that new portions of condensate are removed immediately from the air mass [Dansgaard, 1964]:

$$\delta_p = \frac{\alpha}{\alpha_0} F^{\alpha_m - 1} - 1, \quad (1)$$

where δ (δD or $\delta^{18}O$) is isotope composition expressed as ratio of heavy isotope concentration (mole fraction) in sample to its concentration in standard water, in per mil:

$$\delta = \frac{R_{SA} - R_{ST}}{R_{ST}} \times 1000, \quad (2)$$

R is $\begin{bmatrix} 2H^1H^{16}O \\ 1H_2^{16}O \end{bmatrix}$ or $\begin{bmatrix} 1H_2^{18}O \\ 1H_2^{16}O \end{bmatrix}$, accordingly, α is fractionation coefficient at the given moment of

time, α_0 is the same coefficient in the beginning of the condensation process and α_m is mean α from the beginning to the given moment.

The equilibrium fractionation coefficient α (α_D for deuterium and α_{18} for oxygen 18) is by definition equal to ratio of heavy isotope concentration in liquid to its concentration in water vapor being in equilibrium with the liquid:

$$\alpha = \frac{R_{liquid}}{R_{vapor}} \quad (\text{always} > 1), \quad (3)$$

and can be also determined as ratio of saturation pressure of vapor consisting of light molecules to that of vapor consisting of heavy molecules.

In turn, the fractionation coefficients are temperature-dependent. Experimental determination of α values was made by M. Majoube (1971a):

$$\ln \alpha_{18} = \frac{1137}{T^2} - \frac{0.4156}{T} - \frac{2.0667}{1000} \quad \text{and} \quad (4a)$$

$$\ln \alpha_D = \frac{24844}{T^2} - \frac{76.248}{T} - \frac{52.612}{1000}. \quad (4b)$$

Unlike the condensation process in the atmosphere, the evaporation of water vapor from ocean surface takes place under non-equilibrium conditions. This is due to the fact that water vapor above the sea is under-saturated in respect to water at given temperature. As a consequence, a "kinetic isotope effect" appears due to the slower diffusion of heavy molecules

compared to light molecules, which leads to effective fractionation coefficients being somewhat larger than the equilibrium ones. The first theoretically based model satisfactorily describing the kinetic effect belongs to Merlivat and Jouzel (1979). According to them, the isotope composition of water vapor forming over the sea surface, δ_{v0} , equals to:

$$\delta_{v0} = (1 + \delta_{oc}) \frac{1 - k}{\alpha (1 - kh)} - 1, \quad (5)$$

where δ_{oc} is the isotope composition of the ocean water, α is the equilibrium fractionation coefficient, h is the relative humidity of the air over the water surface and k is the coefficient that accounts for the molecular and turbulent diffusivity of water vapor. The latter coefficient differs by 12 % for D and ^{18}O and is only weakly dependent on wind speed, so in most cases it can be took constant and equal to 0.006 [Johnsen et al., 1989]. One of the most convenient indexes of kinetic effect intensity is "deuterium excess" that is defined as follows [Dansgaard, 1964]:

$$d = \delta\text{D} - 8\delta^{18}\text{O}. \quad (6)$$

Its value is proportional to the effective coefficient of fractionation of water evaporation; this parameter is only slightly changed in the course of equilibrium condensation process, thus carrying quantitative information about the conditions in the moisture source [Jouzel et al., 1982].

The isotope model based on the above equations explains satisfactorily the global relationships between mean annual values of isotope composition of precipitation and air temperature [Dansgaard, 1964]:

$$\delta^{18}\text{O} = 0.7T - 13.6, \quad (7a)$$

$$\delta\text{D} = 5.6T - 100, \quad (7b)$$

as well as between concentrations of δD and $\delta^{18}\text{O}$ in low and middle latitudes [Craig, 1961; Dansgaard, 1964]:

$$\delta\text{D} = 8\delta^{18}\text{O} - 10. \quad (8)$$

However, after measurements of isotope composition in snow samples collected along the inland Antarctic traverses it became clear that the model does not work well for the snow falling in polar regions. Isotope models of Rayleigh type (for equilibrium conditions) gives too high values of deuterium excess thus pointing out a new unknown kinetic effect during the formation of the solid precipitation. This problem was solved in 1984 by Jouzel and Merlivat [Jouzel and Merlivat, 1984] who theoretically explained and empirically confirmed the existence of a kinetic effect during sublimation of water vapor on the surface of ice crystals under the conditions of air supersaturated in water vapor. According to their RMK model (Rayleigh Model taking into account Kinetic effect) the effective coefficient of fractionation in this case equals to:

$$\alpha_e = \alpha \cdot \alpha_k, \quad (9)$$

where α is coefficient of fractionation for water vapor and ice in equilibrium:

$$\ln \alpha_{18} = \frac{11.839}{T} - \frac{28.224}{1000} \quad [\text{Majoube, 1971b}] \text{ and} \quad (10a)$$

$$\ln \alpha_D = \frac{16288}{T^2} - \frac{9.34}{100} \quad [\text{Merlivat, Nief, 1967}], \quad (10b)$$

and α_k is kinetic fractionation coefficient, which is equal to:

$$\alpha_k = \frac{S_i}{1 + \alpha(S_i - 1)D/D'} \quad (\text{always} < 1), \quad (11)$$

where D and D' are the diffusion constants for light and heavy molecules, S_i is supersaturation of air in water vapor with respect to ice.

The largest uncertainty is brought to this model by S_i which is generally not known and is highly difficult to measure (see Section II.1.3). As an optimal solution of this problem the authors suggested to approximate S_i as a function of condensation temperature choosing the coefficient of the function by adjusting the model results to the observed distribution of isotope composition of snow keeping S_i values in reasonable limits [Jouzel, Merlivat, 1984].

The results obtained using RMK are compared in Figure 7 with isotope composition of surface snow samples taken along the traverse Dumont-d'Hurville–Dome C [Lorius and Merlivat, 1977]. It can be seen that the slope of the curve relating $\delta^{18}\text{O}$ in East Antarctic snow with mean annual surface air temperature (T_S) is substantially lower than the slope of the theoretical curve between $\delta^{18}\text{O}$ and temperature of condensation (T_C). This is explained by the fact that the temperature difference at the upper and lower boundaries of surface temperature inversion is increasing while one approaches the center of Antarctica [Connolley, 1996; Phillipot and Zillman, 1970; Voskresensky and Tsigel'nitsky, 1976]. According to Robin (1977), the condensation temperature in Antarctica corresponds to the temperature at the upper boundary of the inversion layer (T_i) within $\pm 4^\circ\text{C}$, i.e.:

$$T_C \approx T_i. \quad (12)$$

Geographical relationship between near-surface and inversion air temperatures obtained using the data from a number of Antarctic sites with T_S range from -15 to -55°C is as follows [Jouzel and Merlivat, 1984]:

$$T_i = C_i T_S - 1.2, \quad (13)$$

where $C_i = 0.67$. If we now draw the relationship between the measured isotope composition of snow and the estimated condensation temperature using equations (12) and (13), then the new curve will coincide exactly with the theoretic one calculated by RMK for the initial sea temperature in the source equal to 20°C (Fig. 7).

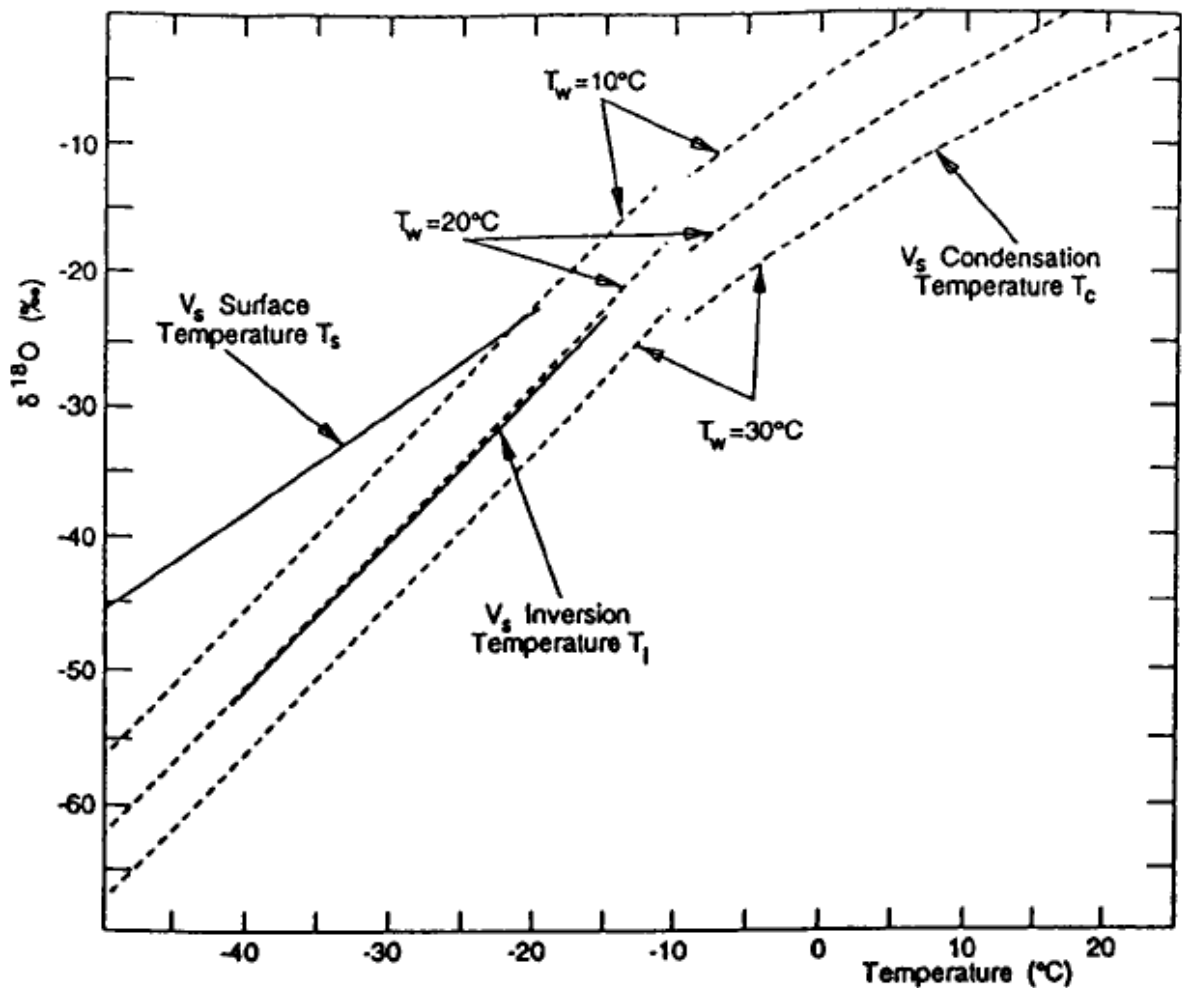


Fig. 7. Changes of $\delta^{18}\text{O}$ concentration in Antarctic precipitation using Rayleigh model (from [Jouzel et al., 1997]). Three sets of curves correspond to different sea surface temperatures (SST) in the source. To calculate isotope composition of liquid and solid precipitation the models of, correspondingly, Merlivat and Jouzel (1979) and of Jouzel and Merlivat (1984) were employed. Solid lines denote changes of surface snow isotope composition along the route Dumont-d'Hurville–Dome C [Lorius, Merlivat, 1977] as a function of surface air temperature (T_s) and inversion temperature (T_i).

One can note the discontinuity of theoretical $\delta(T_c)$ functions for liquid and solid precipitation (Fig. 7). It is due to the fact that at the temperature close to 0°C all the three water phases coexist in clouds [Rogers, 1979], which considerably complicates the isotope transformations during precipitation formation and is not taken into account in the above described models. This lack was filled up in so-called Mixed Cloud Isotope Model (MCIM) [Ciais, Jouzel, 1994]. This required introducing to the model several new parameters, which made the final result more sensible to the model tuning and thus more subjective.

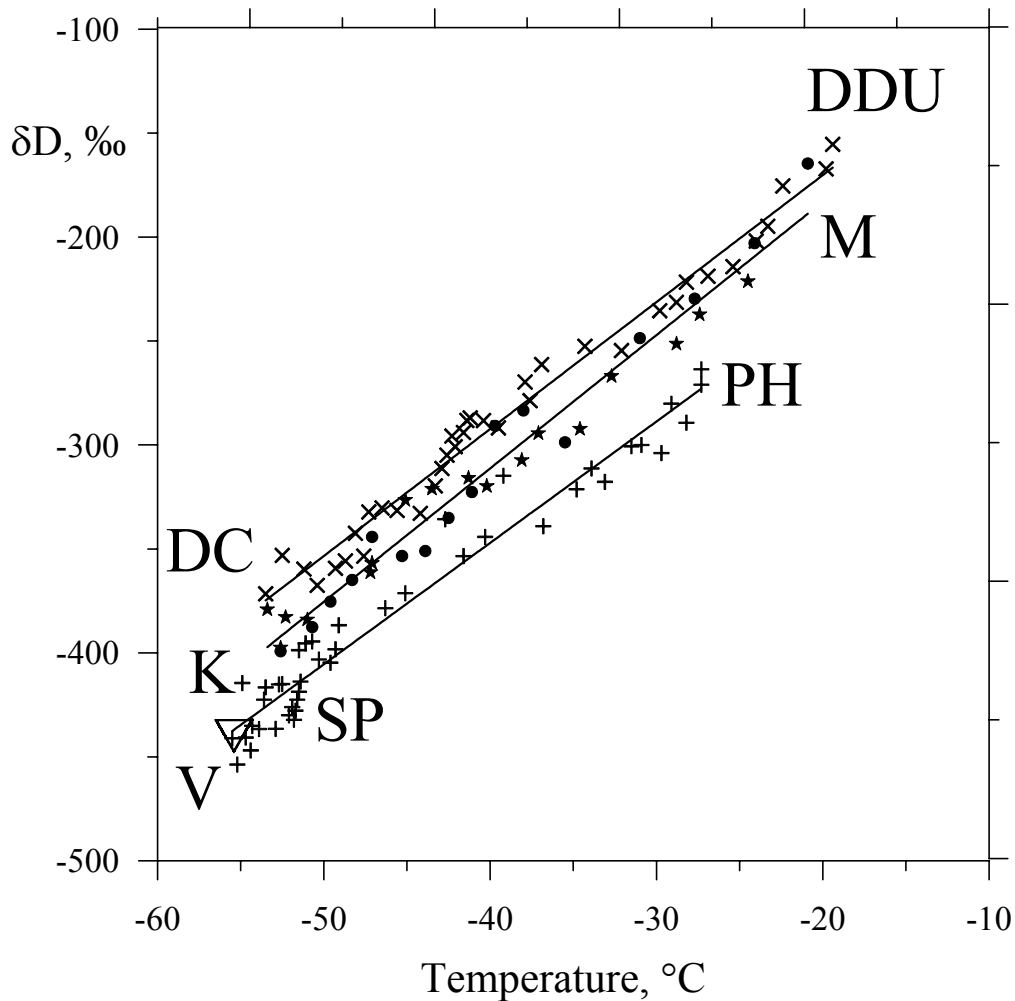


Fig. 8. Relationship between isotope composition of surface snow in Antarctica and 10-m firn temperature (from [Ekaykin et al., 2001] with modifications). Traverses: Dumont-d'Hurville–Dome C [Lorius, Merlivat, 1977]; Mirny–Komsomolskaya [Dahe et al., 1994; Ekaykin et al., 2001]; Patriot Hills–South Pole–Vostok [Dahe et al., 1994]. The inversed triangle denotes Vostok Station characteristics: $\delta D = -440.1 \text{ ‰}$, $T_S = -55.4 \text{ °C}$. The lines represent linear approximations of the traverse data.

In Fig. 8 the relationship between the isotope composition (δD) of surface snow in Antarctica and the 10 m firn temperature[‡] is shown. Despite systematic difference in isotope composition in various sectors of Antarctica, the slope C ($C = d\delta/dT_S = C_T C_i$) in all cases is nearly the same and equals to $6\% \text{ °C}^{-1}$. Taking into account equations (12) and (13), this corresponds to a slope C_T between δD and T_C of $9\% \text{ °C}^{-1}$, which equals to the theoretical C_T coefficient obtained using RMK. Thus, the model is generally in agreement with the experimental data, which validates the model basic assumptions. This encouraged the use of

[‡] which is usually believed to be identical to long-term mean annual near-surface air temperature (T_S). More correctly, 10-m firn temperature is identical to long-term mean annual surface snow temperature which, however, is generally not equal to T_S (see, for example [Mann, Schmidt, 2003]).

present-day geographical relationship between isotope composition of surface snow in Antarctica and mean annual surface air temperature for paleo-temperature interpretation of the isotope profiles from deep ice cores. The first attempt to quantitatively estimate past (over the last 30 ka) changes of surface air temperature using isotope data was made in 1979 by C. Lorius and others (1979). Twenty years after the same approach with only minor corrections was used to reconstruct air temperature changes over the last 420 ka at Vostok Station [Petit et al., 1999]. In particular, in Petit et al. paper (1999) the changes of inversion temperature in the past (compared to its present-day value) ΔT_i is calculated from changes of isotope composition of ice $\Delta\delta D$ according to the following expression:

$$\Delta T_i = \frac{\Delta\delta D - 8\Delta\delta^{18}O_{oc}}{C_T}, \quad (14)$$

where $\Delta\delta^{18}O_{oc}$ is correction for the changes of mean isotope composition of sea water in the past due to changes of water volume trapped in Earth's glaciers, and C_T is regression coefficient taken as $9 \text{ } \text{‰} \text{ } ^\circ\text{C}^{-1}$. In turn, T_i is related to surface air temperature T_S by equation (13). According to this approach, during LGM (Last Glacial Maximum) (δD is about $-483 \text{ } \text{‰}$) T_i and T_S were, respectively, 6 and 8.5°C lower than at present. It was shown later, however, that the sea water isotope composition correction ($\Delta\delta^{18}O_{oc}$) should be taken with the coefficient less than 8 [Jouzel et al., 2003]. It is due to the fact that the influence of isotopic change at the ocean surface weakens as an air mass becomes isotopically depleted, which can be easily demonstrated using a Rayleigh model.

In spite of firm physical basis of "simple" Rayleigh-type isotope models, it can be argued that they do not always adequately reproduce past isotope changes during such global climatic reconstructions as transition from LGM to Holocene [Jouzel et al., 1997]. This is mainly due to the fact that simple models do not take into account changes in atmospheric circulation. This is why since the late 1980s attempts have been made to include isotope transformations during the global water cycle into the General Circulation Models (GCM) [Joussaume et al., 1984].

GCMs calculate temporal changes of various atmospheric fields (wind, temperature, pressure, humidity) in grid points by integrating the equations of motion, continuity, and thermodynamics. Incorporating HDO and H_2^{18}O cycles into GCM involves registration of isotope fractionation (including equilibrium and kinetic effects) during each phase transition, i.e. evaporation from sea surface, condensation of water vapor in the atmosphere, evaporation of drops during precipitation, etc. It has been demonstrated that GCMs satisfactorily reproduce present-day geographical distribution of isotope composition in precipitation and its seasonal variability, as well as spatial relationships between δ and T_S , and δD and $\delta^{18}\text{O}$ (see review in

[Jouzel et al., 1997]). This allowed regarding GCMs to be reliable enough to start modeling of the paleo-climatic atmospheric fields as a complementary method in interpretation of deep ice core data. In general, time scale of such studies is limited to Holocene and LGM because boundary conditions are fairly well known for these epochs, and also because paleodata are available for these periods both for polar and temperate latitudes, which allows verifying the model results. One of the drawbacks of the GCMs is an inability to correctly reproduce processes such as katabatic winds and their influence on surface snow mass balance and isotope composition.

Different types of GCMs give somewhat varying estimations of temporal slope between isotope composition of precipitation and surface air temperature in polar regions. At the same time, it can be noted that in most cases the corresponding C coefficient within 30 % corresponds to present-day geographical slope between δ and T_S (but tends to be less than the latter). On the other hand, model results for high latitudes are rather sensitive to spatial resolution of the model: for central Greenland, temporal C slope for LGM–Holocene transition is $0.89 \text{ ‰ } ^\circ\text{C}^{-1}$ (for $\delta^{18}\text{O}$) for model with $4 \times 5^\circ$ resolution and $0.48 \text{ ‰ } ^\circ\text{C}^{-1}$ for one with $8 \times 10^\circ$ resolution [Jouzel et al., 1994]. GCMs also predict relatively large spatial variability of temporal C slopes, which is typical both for Greenland and Antarctica.

As for East Antarctica where Vostok is situated, rather large range of model estimations of temporal slopes between δ and T_S (for the LGM–Holocene transition) are observed here: some of them exceed the present-day geographical slope [Jouzel et al., 1994], while the others are significantly less than it [Noon, Simmonds, 2001].

In general, this brief review shows that in spite of great potential of GCM application in paleoclimatic studies, one should be cautious when using the model results of such complicated processes as change in isotope composition of polar precipitation during climatic transition from glacial age to inter-glacial epoch.

1.2.2. Empirical estimations of relationship between isotope composition of precipitation and temperature

The use of the isotope signal as paleo-thermometer is based on the assumption that the present-day geographic (spatial) slope between δ in precipitation and T_S is equal to the corresponding temporal slope. This assumption needs additional empirical validation because the isotope composition of precipitation in the past could be influenced by other factors than condensation temperature, e.g., changes of evaporation conditions in moisture source and seasonality of precipitation (see review in [Jouzel et al., 1997, 2003]). Moreover, independent

estimates of past ice sheet surface temperature based on borehole thermometry show that the above described isotope approach underestimates the amplitude of temperature changes in Greenland by factor of two [Cuffey et al., 1995; Johnsen et al., 1995] and in Antarctica by about 30 % [Salamatin et al., 1998a, b].

Direct comparison of isotope composition and air temperature. In terms of experimental validation of δ/T_S relationship the most attractive approach is direct comparison of recent snow isotope composition with instrumental temperature measurements carried out at the same site during a sufficiently long period of time. Only two polar stations meet these conditions: South Pole and Vostok. At South Pole a significant linear relationship was found between mean annual snow isotope composition (δD) values as measured on samples from pits and mean annual surface air temperature, the coefficient of regression being $20 \text{ ‰ } ^\circ\text{C}^{-1}$ [Jouzel et al., 1983]. Such a large value of C (three times larger than corresponding present-day geographical slope) is still not very well understood, but could be explained either by difference between precipitation-weighted mean and simple mean air temperature, or by difference between temperature variability at the surface level and at the level of condensation.

As for Vostok Station, prior to the beginning of this study three papers have been published concerning isotope composition of precipitation and deposited snow. Gordienko et al. (1976) presented data on seasonal changes of isotope composition ($\delta^{18}\text{O}$) of snow precipitation at Vostok in 1970. They obtained linear relation between monthly values of δ and surface air temperature with the slope of $0.84 \text{ ‰ } ^\circ\text{C}^{-1}$ that corresponds to the slope of $6.7 \text{ ‰ } ^\circ\text{C}^{-1}$ for δD . This value agrees well with the theoretical slope from simple isotope model. In papers of Dansgaard et al. (1977) and Kolokolov et al. (1993) the isotope profiles are shown from two 1.2-m deep snow pits dug in 1975 and 1981. Mean $\delta^{18}\text{O}$ values of snow deposits are, correspondingly, -57.9 and -56.3 ‰ . No comparison of isotope data with temperature records was made.

In the other parts of Antarctica and in Greenland the comparison of δ and T is limited either by too short series of temperature observations or by too large distances between sites of isotope study and sites of temperature measurements. Nevertheless, in all cases the slope of temporal relationship between changes of δ and T_S was found to be considerably (up to 40–50 %) less than corresponding present-day geographical slope (see review in [Jouzel et al., 1997]). The observed difference can be real or might be related to comparatively low correlation coefficients between the series either due to "stratigraphic" noise in the isotope composition records or due to long distance between sites.

Borehole thermometry. Among the methods of independent past temperature estimations in polar areas, the measurement of deep borehole temperature profiles provides information on temperature changes over the last centuries and millennia. The largest uncertainty of the method is related to the fact that observed temperature profiles could be formed as a result of various climatic histories. Consequently, to reconstruct past changes of T_S one should a priori assume constraints of possible scenarios of temperature variations[§]. The most effective approach is to put the isotope profile into the thermo-dynamic model of ice sheet as an approximation of past temperature changes. Then by tuning the C coefficient between past changes of isotope composition and surface ice temperature one can find the best correspondence between the observed borehole temperature profile and that produced by the model. This method allows answering the question whether the isotope profile is good paleo-thermometer or not and at the same time to calibrate this thermometer. The results obtained in Greenland and Antarctica show that the isotope composition of ice deposits, indeed, can be interpreted in terms of past temperature changes. The second conclusion is that the calibration coefficient C is not constant in time: for Holocene it is close to the present-day slope of geographical δ/T relationship, while for the LGM temporal C is roughly 2 times for Greenland [Cuffey et al., 1995; Johnsen et al., 1995] and by 30 % for Antarctica [Salamatin et al., 1998a and b] less than spatial C .

Use of melt layers. To estimate past changes of summer temperature in polar regions one can use the percentage of melt layers in ice cores. For instance, Koerner and Fisher (1990) used this method to estimate temperature changes on the Agassiz ice cap for the period of 8.5 ka. They concluded that during this period the summer temperature decreased on average by 2 °C, which corresponds well to the isotope-temperature reconstruction carried out using the present-day spatial C slope. The same conclusion was drawn by Alley and Anandakrishnan (1995) who studied the occurrence of melt layers in middle and late Holocene on ice core from GISP2 borehole. The main drawback of this method is the fact that the number of days with temperature exceeding 0 °C is related not only with mean summer temperature, but also with variability of mean daily temperatures in summer period. Given the same mean summer temperature, the site where variability (standard deviation) of mean daily temperatures is higher would possess more frequent melt layers in the ice thickness.

Correlation with snow accumulation rate. Estimations of past temperature changes can be performed by correlation of isotope profile with accumulation series [Ritz, 1992]. It is supposed that the snow accumulation rate is a function of saturation water vapor pressure which in turn is

[§] Another difficulty is related with above mentioned fact that ice sheet surface temperature recorded in a borehole thermogram does not equal to T_S . The discussion of this issue is beyond the scope of the present work

temperature dependent. If past accumulation changes are known (from annual layer thickness measurements in Greenland or by measurements of ^{10}Be concentration in Antarctica) than one can choose such coefficient C between isotopes and temperature that calculated accumulation series will be consistent with the observed one. In particular, for Greenland this method confirmed in general that temporal C slope is less than corresponding spatial slope [Jouzel et al., 1997]. This method does not properly take into account, however, the fact that snow accumulation rate is not only a function of air saturation by water vapor but also of atmospheric circulation. For instance, the decrease of accumulation rate in Greenland in LGM is explained to a large extent by changes of cyclones paths [Kapsner et al., 1995]. In case of central Antarctica the situation is even more uncertain because of relatively little knowledge of past atmospheric circulation in this region.

A more sophisticated method was developed by F. Parrenin with co-authors for Vostok Station [Parrenin et al., 2001]. Its main assumption is that the ice core may be dated by correlation of precession cycles of solar insolation (about 20 kyrs) with the profile of isotope composition ($\delta^{18}\text{O}$) of oxygen trapped in the ice. Then, the isotope composition of the ice is supposed to be related to condensation temperature by a second-order equation, while relationship between accumulation rate and T_C is taken as in the previous method. The obtained results suggest that "classical" isotope method underestimate amplitude of past temperature changes by 0–30%.

Data on gas inclusions. Additional information on past snow accumulation rate and ice sheet surface temperature come from experimental data on number and size of gas inclusions in ice. It was established that geometrical characteristics of hypogene inclusions are determined by absolute dimensions of ice grains at the pore close-off depth. The dimensions are in turn a complex function of initial surface snow density, as well as of mean (over the period of firn formation) firn temperature and snow accumulation rate [Lipenkov, 2000]. A model was developed to calculate ratio between snow accumulation rate and surface snow temperature in the past using the data on mean air bubble amount in ice. In particular, for Vostok this method suggests more severe (i.e., lower values of temperature and snow accumulation rate) conditions during LGM than according to classical isotope approach using present-day spatial coefficient C .

Isotope composition of trapped air. Recently, the method developed by J. Severinghaus et al. (1998) was used for estimation of paleo-temperature at Vostok Station. The approach is based on the measurements of isotope composition of air gases (first of all, nitrogen and argon) trapped in the ice. During a rapid snow surface temperature change, a firn vertical temperature gradient is induced, which leads to thermal fractionation of gases. After pore close-off the isotope

composition of gases in the inclusions does not change and thus their measurement allow reconstructing the rapid surface temperature changes when the air was trapped. This method was successfully used in Greenland to estimate the temperature jump in the end of Younger Dryas [Severinghaus et al., 1998] and showed larger amplitudes of temperature change than using spatial isotope-temperature relationship. The use of this method in Antarctica is difficult due to the fact that the climate of this continent during the last 400 ka was not subjected to such rapid changes as in Greenland [Petit et al., 1999]. Nevertheless, N. Caillon et al. (2001) made an attempt to estimate the rapid temperature rise at the end of cold isotope Stage 5.4 (107 kyrs BP) when the intensity of temperature changes was the highest. They concluded that the amplitude of warming was by $20\pm 15\%$ larger than according to "classical" interpretation (equation (14)).

As a whole, the review of the results of experimental studies suggests that present-day spatial δ/T_S relationship cannot be a priori used for interpreting the isotope data for the period prior to Holocene. This conclusion requires deeper insight into the factors affecting the relationship between isotope composition of polar precipitation and air temperature in precipitation site.

I.3. Factors influencing the relationship between snow isotope composition and surface air temperature

In general, the factors influencing the relationship between δ and T_S can be divided into two categories: the first is related to the fact that other parameters than condensation temperature do participate to the formation of isotope composition of snow thickness (moisture source conditions, microphysical processes in clouds, snow redistribution by wind and post-depositional processes). The second group is related to biases in the δ/T_S ratio because of discontinuity of the precipitation and due to difference between condensation temperature and T_S [Jouzel et al., 1997].

I.3.1. Moisture source conditions.

As mentioned in Section I.2.1, the isotope composition of precipitation is related, strictly speaking, to the difference between temperature of condensation and temperature in the source, rather than to the condensation temperature itself. Thus, the common practice of using present-day relationship between δ and air temperature in the site of snow sampling in paleoclimatic reconstructions suggests constant temperature in the moisture source in the past, which is generally not true.

It can be easily shown that cooling in the moisture source with a simultaneous decrease of condensation temperature in Antarctica will result in a lower apparent C slope ($d\delta/dT_S$). In particular, to explain this phenomenon, the following simplified scheme can be proposed. The lower is the ocean surface temperature the heavier is the isotope composition of precipitation given the same condensation temperature because of less isotope depletion (equation (1)). For example, if Sea Surface Temperature decreases by ΔSST , the isotope composition of precipitation in Antarctica will increase by $C\Delta SST\%$ (here C is spatial coefficient which is supposed to be constant in time). If condensation temperature at the same time decreases by ΔT_C , than the total change of isotope composition would be $(-C\Delta T_C + C\Delta SST)$, and the temporal slope between δ and T_C changes would equal to $\frac{-C\Delta T_C + C\Delta SST}{-\Delta T_C} = C\left(1 - \frac{\Delta SST}{\Delta T_C}\right) < C$. From

this equation it can be seen that if source and condensation temperatures changed by the same value, the isotope composition would not change at all. According to Boyle (1997), this effect alone is able to explain the observed difference of spatial and temporal C slopes with ΔSST equal to only 5°C . The additional difference between the slopes is brought by increasing of sea water isotope composition during LGM.

In reality, the above scheme is too simplified since SST change also causes the change in spatial C slope.

This idea was developed by K. Cuffey, F. Vimeux and J. Jouzel [Cuffey, Vimeux, 2001; Vimeux et al., 2002]. They suggested using the deuterium excess signal (equation (6)) measured in ice core samples for moisture source-temperature correction of paleo-temperature reconstructions. As mentioned above, deuterium excess (d) carries information about conditions of evaporation from sea surface and, in particular, about sea surface temperature. However, the Rayleigh model also suggests that d depends on condensation temperature T_C (through slight changes of C slope) and on isotope composition of sea water (δ_{oc}). In turn, δD of ice is also related to all these three parameters, which allows setting up the following equations:

$$\Delta\delta D = \gamma_C\Delta T_C - \gamma_{SST}\Delta SST + \gamma_{oc}\Delta\delta_{oc} \quad (15a)$$

$$\Delta d = -\beta_C\Delta T_C + \beta_{SST}\Delta SST - \beta_{oc}\Delta\delta_{oc}. \quad (15b)$$

where all the γ and β coefficients are > 0 . Solution of these equations gives the following expressions for the condensation and sea surface temperature:

$$\Delta SST = (\gamma_C/\Omega)\Delta d + (\beta_C/\Omega)\Delta\delta D + ((\gamma_C\beta_{oc} - \beta_C\gamma_{oc})/\Omega)\Delta\delta_{oc} \quad (16a)$$

$$\Delta T_C = (\gamma_{SST}/\Omega)\Delta d + (\beta_{SST}/\Omega)\Delta\delta D + ((\gamma_{SST}\beta_{oc} - \beta_{SST}\gamma_{oc})/\Omega)\Delta\delta_{oc}, \quad (16b)$$

$$\Omega = \gamma_C\beta_{SST} - \gamma_{SST}\beta_C. \quad (16c)$$

Coefficients γ and β are determined from Rayleigh isotope model. Their values, as well as the sensibility of ΔSST and ΔT_C to variability of each of the coefficients, are listed in Table 2.

The use of this approach for interpretation of isotope data from Vostok ice core showed that correction on source temperature change does not exceed 1°C and thus effective temporal slope C for LGM–Holocene transition is less than the present-day geographical slope by only about 15 %.

The main drawback of this method is the fact that deuterium excess is highly sensible to air supersaturation (S_i) by water vapor over Antarctica. Past changes of S_i are not known and likely are not properly taken into account in this approach. Besides, this model does not consider past changes of other source parameters (humidity and coefficient of turbulent diffusion, see equation (5)). Partly, humidity is accounted for by its relation to sea surface temperature [Vimeux et al., 2001]. Besides, GCM experiments shows that past variability of air humidity was relatively low [Bush, Philander, 1999]. As for turbulent regime of atmosphere, it is also generally assumed that this parameter did not change a lot in the past [Johnsen et al., 1989].

Table 2. Coefficients of equations (15)–(16) and sensibility of temperature to their variability (from [Vimeux et al., 2002]).

Coefficient	Definition	Value	Units	$\sigma(\Delta T_C)$, $^\circ\text{C}$	$\sigma(\Delta SST)$, $^\circ\text{C}$
γ_C	$d\delta D/dT_C$	7.1	$\text{‰ }^\circ\text{C}^{-1}$	1	0.4
γ_{SST}	$d\delta D/dSST$	3.7	$\text{‰ }^\circ\text{C}^{-1}$	0.3	0.1
γ_{oc}	$d\delta D/d\delta_{oc}$	4.8	–	0.1	0.05
β_C	$ddxs/dT_C$	0.5	$\text{‰ }^\circ\text{C}^{-1}$	0.2	0.4
β_{SST}	$ddxs/dSST$	1.3	$\text{‰ }^\circ\text{C}^{-1}$	0.3	0.6
β_{oc}	$ddxs/d\delta_{oc}$	2.8	–	0.2	0.3

Sensibility of ΔSST and ΔT_C for LGM–Holocene transition was determined by Monte-Carlo method for the variability (1 σ of normal distribution) of each coefficient equal to 10 % of its value.

It should be noted that changes of source conditions can be due to both 1) temporal changes of atmospheric and ocean parameters of the basins from where water vapor comes to given polar site, and 2) geographical shift (mainly, latitudinal) of those basins, as well as by changes of relative contribution of each basin into the total amount of precipitation at the site of study. The latter can be illustrated by Figure 8 which demonstrates that the difference of isotope composition between different Antarctic sectors can reach about 50 % (for δD). This value is the

same order of magnitude as change of ice isotope composition at Vostok during LGM–Holocene transition. Besides, coastal areas of Greenland and Antarctica experienced substantial changes of sea ice extent in the past, which could strongly influence the contribution of Antarctic seas into the precipitation amount in these areas.

I.3.2. Seasonality of precipitation.

As noted by E. Steig with co-authors [Steig et al., 1994], the inter-annual variability of any parameter of snow thickness (accumulation rate, chemical or isotope composition) that has comparatively large seasonal variations would be sensitive to changes of intra-annual distribution of this parameter. It means that the isotope composition of snow is related to mean air temperature weighted by amount of precipitation, rather than to simple mean air temperature. For example, if most of the precipitation falls during the cold period (as it is the case for Vostok), the mean weighted surface air temperature is less than T_S . If in the past the contribution of winter precipitation was lower, the mean isotope composition of ice deposits would be higher at the same mean air temperature. In particular, the discrepancy between paleo-temperature reconstruction based on classical isotope method and on borehole thermometry in Greenland are presently explained by a decrease of winter precipitation amount due to stronger zonal circulation during LGM [Werner et al., 2000]. Similar mechanism could be assumed for Antarctica (especially for the coastal parts of the continent) but there are no reliable data on the past changes on precipitation seasonality there.

Another side of this problem is the fact that the weather during days with and without precipitation is different: in the former case (days with precipitation) air pressure is lower, while temperature, wind speed and humidity are higher (see Section I.1.3). This results in "isotope" temperature being systematically higher than average annual temperature. Taking into account that annual accumulation layer in central Antarctica can be formed as a result of few snowfalls, then it becomes clear that mean annual air temperature is in general not representative in terms of conditions of precipitation formation in this area. However, the results of Section III.1 of this work suggest that most of precipitation at Vostok consists of ice crystals from clear sky. In this case isotope composition of snow is well characterized by mean annual air temperature at the condensation level.

I.3.3. Microphysical conditions of precipitation formation.

The largest uncertainty of all the simple isotope models is related to the parameter of air supersaturation by water vapor that controls kinetic fractionation during snow formation in polar area (equation (11)). First of all, model results are sensitive to the value of temperature below

which there is no more liquid droplets in the atmosphere, i.e., the temperature of commencement of non-equilibrium process of ice crystals growth. In turn, D. Fisher [Fisher, 1991] pointed out that this critical temperature is related to the impurity concentration in the atmosphere. The temperature–supersaturation function itself depends on concentration of condensation nuclei, too. So, during the glacial epochs when dust content in the air was much higher than now [Petit et al., 1999], precipitation formation could take place at lower values of supersaturation S_i , which would lead to lower values of isotope composition of ice. Thus, neglect of this factor leads to an overestimation of the amplitude of past temperature changes, which to some extent compensates the influence of two previous factors.

It should be noted that even disregarding possible variations of air supersaturation in the past due to impurities concentration changes, it is not justified to use the present-day S_i -temperature relationship (which coefficients were adjusted to reproduce modern geographical distribution of δ and d) for interpretation of paleo-data. It is mainly related to the fact that adjustment of the S_i function by spatial distribution of T_S , δ and d in Antarctica suggests, in fact, that precipitation in central and coastal parts of the continent come from the same source, which is not true. Indeed, the coastal areas are under strong influence of local conditions (e.g., [Masson-Delmotte, in press]), while inland parts of Antarctica receive moisture from tropics and sub-tropics [Vimeux et al., 2001]. Thus, change of snow isotope composition as one goes from coast to the center of the continent is determined not only by successive distillation of water vapor in a given water mass, but also by diminishing of the contribution of high-latitude sources in total precipitation amount.

I.3.4. Difference between condensation and surface air temperature.

Strictly speaking, the isotope composition of precipitation is related to the temperature of atmosphere at the level of condensation (T_C) and not with surface air temperature (T_S). Consequently, to interpret isotope data in terms of surface air temperature one must know the relationship between T_S and T_C .

According to G. Robin (1977), condensation temperature is equal, with some uncertainty, to the temperature at the upper boundary of surface inversion layer T_i (see equation (12)). Beyond this idea is the fact that the upper part of the inversion represents the warmest and the most water-saturated layer of atmosphere, and its cooling leads to moisture sublimation at the temperature close to (or slightly lower than) T_i . Although it seems obvious, this scheme looks too simplified because in central Antarctica there are three types of precipitation each having its own

condensation temperature [Averianov, 1990]: precipitation from clouds, from clear sky (ice crystals) and rime (the contribution of the latter is negligible, as mentioned in Section I.1.3).

It should be noted that even if condensation temperatures were strictly equal to inversion temperatures in central Antarctica, using present-day spatial relationship between T_i and T_S (equation (13)) as an approximation of temporal relation between T_C and T_S is incorrect. The difference between T_i and T_C is increasing linearly from about zero in central Antarctica to the maximum value at the coast, where inversion temperature practically equals to T_S while condensation temperature is considerably colder than T_S since the precipitating clouds are at the height of 1–2 km above the surface [Averianov, 1990] (Figure 9). Thus, the spatial slope $C_i = dT_i/dT_S$ is larger than the spatial slope $C_C = dT_C/dT_S$. For example, for Mirny Station T_S , T_i and T_C equal to, correspondingly, -11, -12 and -15 – -18°C (for the condensation level 1000–2000 m, see Section I.1.3) [Aerologicheskiy spravochnik Antarktity, 1967; Averianov, 1990]. Thus, the mean spatial gradients C_i and C_C for the Mirny–Vostok profile are equal, correspondingly, to 0.60 and 0.52–0.45. It means that the dT_i/dT_S slope of 0.67 used in the simple isotope models is overestimated by 25–30 %. The fact that models still succeed to reproduce the present-day spatial relationships between snow isotope composition and surface air temperature should not deceive, because, as mentioned above, the model is adjusted to the experimental data through the tuning of the S_i parameter. So, the real C_C slope must be somewhat less than 9 ‰ °C⁻¹ simulated by the model.

We argue that neither present-day spatial, nor modern (over the period of instrumental temperature measurements in Antarctica) temporal (inter- or intra-annual) relationships between condensation and surface air temperature cannot be a priori used for interpretation of paleoclimatological data (in particular, for the LGM–Holocene transition). This is mainly due to the fact that the intensity of inversion depends on the radiation regime of the ice sheet surface that changed considerably in the past. Since variations of incoming solar radiation (and thus probably of surface radiation balance) do not parallel to temperature variations in the time scale of interest [Petit et al., 1999], temporal coefficients C_i and C_C are likely not constant. This is indirectly confirmed by the conclusion made by A. Salamatin et al. (1998) that additional signal with a precessional period of ~20 kyrs (one of the main insolation periods) must be introduced into the series of inversion temperature in order to interpret isotope profile in terms of T_S .

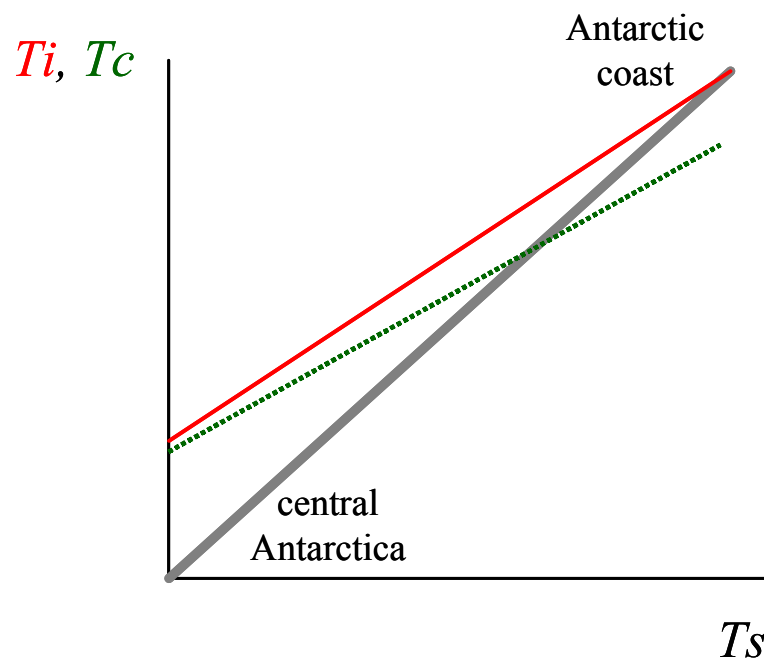


Fig. 9. Sketch illustrating the difference between spatial relationships of dT_i/dT_S and dT_C/dT_S slopes. The thick gray line is the line of equal values, the thin solid and dotted lines are spatial distribution of, correspondingly, inversion (T_i) and condensation (T_C) temperatures.

I.3.5. Glaciological factors.

Past air temperature change could take place not only because of climatic variations, but also due to change of ice sheet thickness and of its height over the sea surface. The role of this factor must be the strongest for the coastal areas of the ice sheet where variations of the thickness were the largest. Changes of precipitation isotope composition in this case must have been related to changes of condensation temperature, but also, to less extent, to changes of air pressure at the level of condensation. As for the area of Vostok, both model simulations [Salamatin et al., 1998] and experimental data [Lipenkov, 2000] show that change of ice sheet surface height during the LGM–Holocene transition was about 130 m. This implies that temperature shift due to the glaciological factor could be of the order of 1 °C, or roughly 5–10 % of the corresponding climatic shift.

I.3.6. Post-depositional processes.

Up to now we considered processes influencing the relationship between air temperature and isotope composition of precipitation. One should not forget, however, that the object of our study is snow or ice deposits for which the isotope composition can be different from that of precipitation due to the processes of deposition, re-deposition and snow metamorphism.

Wind activity plays a large role in the redistribution of freshly fallen snow resulting in its irregular deposition (e.g., [Black and Budd, 1964; Ekaykin et al., 1998; Gow and Rowland, 1965; Van der Veen et al., 1999]). This process is likely important for the isotope composition, too (e.g., seasonal wind erosion, i.e., preferential removal of snow deposits of a given season [Fisher et al., 1985]), though this issue was not sufficiently studied up to now.

After final deposition of fresh snow at the snow surface a number of post-deposition processes begin which change its initial physical and chemical properties. At Vostok, with its extremely low snow accumulation rate (about 7 cm of snow per year), a new snow layer spends a long time near the surface where temperature gradients, and thus strength of metamorphism processes, are maximal. As a result, initial properties of precipitation can be considerably altered. Concerning the isotope composition of the snow, the most important process is likely summer snow sublimation and formation of depth hoar layers. The isotope composition of depth hoar is, as a rule, by few permil higher than that of ambient snow [Ekaykin et al., 2002] due to vapor exchange with snow grains [Waddington et al., 2002] and, probably, to isotope fractionation during the process of sublimation [Golubev et al., 2000]. Since the strength of post-depositional effect is inversely related to the snow accumulation rate [Waddington et al., 2002], one could guess that during LGM, when accumulation was much less than now [Petit et al., 1999], difference between δ of ice deposits and that of precipitation was larger. Thus, according to this factor the amplitude of temperature change during LGM–Holocene transition is probably underestimated. Our preliminary estimations (see Sector IV.3) show that the corresponding correction might be as large as -12‰ , or about 25–30 % of the total δD shift during the LGM–Holocene transition. In this case, the post-depositional effect alone could explain the observed discrepancy between the "classical" paleo-temperature reconstructions (based on the interpretation of the deep ice core isotope profile) and those based on the borehole thermometry.

Finally, just after the snow deposition the isotope variations start to smooth out as a result of water vapor diffusion in the firn pores [Johnsen, 1977; Johnsen et al., 2000]. Since this process does not change the mean isotope composition value for sufficiently long (an order of first tens of centimeters) isotope profiles, it does not influence the observed relation between δ and T .

As a conclusion of the above presented review of the main factors influencing the relationship between changes of isotope composition of ice deposits ($\Delta\delta_{\text{ice}}$) and changes of mean annual surface air temperature (ΔT_S), we propose the following equation relating these two parameters (compare to equations (14) and (16)):

$$\Delta T_S = \frac{(\Delta \delta_{ice} + \Delta d\delta) + C_{SST} \Delta SST - C_{oc} \Delta \delta_{oc}}{(C_T + \Delta C_T) C_{seas} C_C}, \quad (17)$$

where Δ denotes difference of a parameter value between past and present;

$d\delta$ – change of isotope composition of precipitation during post-depositional processes ($d\delta = \delta_{precip} - \delta_{ice}$); for LGM $\Delta d\delta$ is unknown (it is likely negative and can be related to the variations of snow accumulation rate);

C_{SST} accounts for change of isotope composition of precipitation as a result of changes of moisture source conditions ($C_{SST} = \Delta \delta_{precip} / \Delta SST$), $C_{SST} > 0$, and it can be determined from simple isotope model;

C_{oc} accounts for change of isotope composition of precipitation due to change of sea water isotope composition ($C_{oc} = \Delta \delta_{precip} / \Delta \delta_{oc}$), $C_{oc} > 0$, and it can be determined from simple isotope model;

C_T accounts for the relationship between isotope composition of precipitation and mean weighted condensation temperature ($C_T = \Delta \delta_{precip} / \Delta T_C^w$) for present-day conditions ($C_T > 0$ and also includes relationship between δ_{precip} and air supersaturation S_i ($S_i = f(T_C^w)$). If the function $S_i = f(T_C^w)$ is known, C_T can be determined from simple isotope model). ΔC_T accounts for past changes of C_T related mainly to temporal changes of the function $S_i = f(T_C^w)$. For LGM ΔC_T is likely positive;

C_{seas} accounts for precipitation seasonality changes in the past ($C_{seas} = \Delta T_C^w / \Delta T_C$, where T_C is mean annual air temperature at the level of condensation); $C_{seas} > 0$ and for Vostok Station it is unknown (it can be larger than unity, which means more contribution of winter precipitation in the past, or smaller than unity, which implies larger contribution of summer precipitation);

C_C accounts for the relationship between mean annual air temperature at the condensation level and surface mean annual air temperature ($C_C = \Delta T_C / \Delta T_S$), $C_C > 0$ and is, likely, not constant in time (can be a function of surface radiation balance).

Most coefficients in equation (17) are not known, which implies that one should be cautious with the results of isotope-temperature reconstructions made on the base of classical approach using present-day spatial δ/T_S relationship. Nevertheless, a number of experimental data and GCM simulations (see the above review) show that the difference between spatial and temporal δ/T_S slopes likely does not exceed 30 %. This allowed the statement that present day spatial C slope is equal, within the uncertainty of its value, to the corresponding temporal slope and thus can be used as a surrogate of the latter [Jouzel et al., 1997; 2003]. The fact that coefficients in equation (17) have opposite signs does not rule out the possibility that above described factors counterbalance each other, making the mentioned slopes equal. To answer this

question a number of special studies (e.g., study of past changes of the source conditions, investigation of microphysical processes during the snow formation and post-depositional processes in snow thickness, etc.) are needed that are far beyond the scope of this work.

I.4. Conclusion of Chapter I

In conclusion of the presented review of literature we should note that the modern meteorological regime of central Antarctica and its intra- and inter-annual variability are fairly well documented. On the other hand, the relationship between local climate and the main circulation regimes of Southern Hemisphere is not investigated in sufficient details. The meteorological conditions of precipitation formation in central Antarctica, as well as precipitation rate and its temporal variability are not accurately known. As for present-day isotope composition of precipitating and deposited snow at Vostok, this topic was almost not touched upon. The role of various factors in formation of isotope composition of snow thickness (in particular, snow redistribution by wind) has not been completely discussed as well. Thus, the above review validates the necessity and significance of the present work.

II. METHODS AND EXPERIMENTAL DATA

II.1. Experimental data

II.1.1. Meteorological data.

Meteorological observations have been carried out at Vostok since December 1957 with interruptions in 1962, 1982, 1994, 1996 and 2003. The readings are made 4 times a day (1 a.m., 7 a.m., 1 p.m. and 7 p.m. local time) and include measurements of air and surface snow temperature, wind speed and direction, air pressure, humidity, precipitation, cloudiness and meteor. To carry out the present study, we created a dataset of mean 10-day, monthly and yearly values of these characteristics. The source of the data is Reference book on Antarctic climate (1977), as well as archive materials (meteorological tables TM-1) of Arctic and Antarctic Research Institute (AARI). The data are stored on a personal computer as files of "Excel" format. Mean monthly values of several meteorological characteristics for Vostok and other Russian Antarctic stations are available at the official site of AARI:

http://www.aari.nw.ru/projects/Antarctic/default_en.asp

Also, during the period from February to December 2000 meteorologist of Vostok Station V.A. Persky collected (specially for this study) samples of precipitation and blowing snow to be measured for isotope composition (see Section II.2.3), and for each sample he registered meteorological data for the period of sample accumulation.

As pointed out in Section I.1.3, data on precipitation amount are not reliable due to wind bias. Also, measurements of air humidity at low temperature are a very difficult task [Burova et al., 1990], so these data are also considered as unreliable.

In our work (Section III.1) we also used the data on the days with the cloud precipitation. In many cases the "cloud" precipitation was detected when no lower or middle clouds were observed (only the upper ones). In these cases we considered these precipitation as "diamond dust" (i.e., precipitation from clear sky), following the definition of this type of precipitation and keeping in mind that the upper clouds (C_i , C_s and C_c) cannot give "normal" precipitation ("snow" or "snow grains", in contrast to "diamond dust") [Khromov, Petrosyants, 2001].

II.1.2. Balloon-sounding data.

Aerological (balloon-sounding) observations were carried out at Vostok from December 1957 until January 1992 with interruptions in 1962 and 1982. Balloons were launched once a day at 7 a.m. local time (sometimes twice or even four times a day to study diurnal variations of

atmospheric parameters). The measured characteristics were temperature, pressure, air humidity, wind and lower cloud boundary. For the present study, we chose data at the isobaric surfaces of 600, 500 and 400 hPa (these data characterize the atmospheric layer where the bulk of precipitation is formed), as well as at the boundaries of surface inversion and isothermal layers, for the whole period of observations. The data set includes daily data, which were then used to calculate mean 10-day, monthly and yearly values. The source of the data is AARI's archive (aerological tables TAE-7 for period 1960–1963 and TAE-16 for period 1964–1991), as well as published data for 1958 [Tret'ya KAE, 1962] and 1959 [Chetvertaya KAE, 1963]. Besides, mean monthly data of main meteorological parameters for the first years of observations are published in the Reference book on aerological data from Antarctica [Aerologicheskii spravochnik Antarktidiy, 1967]. The data set was created during the period from 2000 to 2003 with intensive participation of students of the Faculty of Geography of St. Petersburg State University A. Bykova, D. Anufrieva, A. Zakharov and Yu. Popov, as well as of a collaborator of Department of Geography of Polar Regions of AARI A.V. Preobrazhenskaya. It should be underlined that by the beginning of this study there was no available electronic database on balloon-sounding data from Vostok, and up to now there is no published work summarizing aerological data over the whole period of observations. Nevertheless, in nearest future mean monthly balloon-sounding data from Vostok and other Russian Antarctic stations shall be put on the official web-site of AARI (www.aari.nw.ru).

According to personal communications of V.E. Lagun and N.N. Kazakova (2003), balloon-sounding series from Vostok may contain methodological and statistical inhomogeneities related to the changes of radiosonde types used, as well as to professional skill of observers. In relation to this, we investigated the statistical homogeneity (of mean values and variance) [Panovsky and Brayer, 1972] of the series of mean annual values of main meteorological parameters at the isobaric surfaces and at the upper boundary of surface inversion layer. In each series breaks of homogeneity were found with a tendency to group around the same years. Most often, homogeneity broke in 1962 (both for mean and variance), which can be seen most clearly in series of inversion characteristics: in 1958–1961 inversions were thicker but less intensive and less frequent than during the successive period. Also, homogeneity broke around 1969 which can be seen in series of pressure and temperature (for mean values) and wind speed and humidity (for variance), and around 1980 in series of pressure (for mean values) and humidity (for variance).

At least in one case (1962) the break of homogeneity coincides with change of radiosonde type (see Table 3).

According to Kazakova (1986) and Zaitseva (1991), the "P3-049" system had significant deficiencies in measurements of temperature and pressure. In particular, the temperature sensor was comparatively inert, which explains the above-mentioned inhomogeneity of inversion characteristics. Due to this large inertia, the radiosonde was not able to fix the maximum temperature of the inversion layer, which led to low observed inversion intensity (or even omitted the inversion layer completely, which cause its low observed frequency). On the other hand, the radiosonde "remembered" high temperatures when the inversion layer had already been passed, which explains the comparatively large thickness of inversion layer. Thus, this led us to the decision of rejecting the balloon-sounding data for the period of 1958–1961. It should be noted that using these data could result in observed false trends in studied series. Probably, this fact explains mentioned in literature trends of inversion characteristics over 1958–1982 [Tsigel'nitsky, 1990] (see Section I.1.2).

Table 3. Systems of radiosondes used in balloon-sounding observations at Vostok Station (based on data from [Kazakova, 1986; Maystrova, Kazakova, 1991]).

Radiosonde type	"P3-049"	"Malakhit-A-22"	"Meteorit" PK3-5	"MAP3 2-2"
Period of exploitation	Dec 1957 – Dec 1961	Jan 1962 – Sept 1977	Oct 1977 – Jan 1987	Feb 1987 – Jan 1992

As for the breaks of homogeneity in 1969 and 1980, they do not correspond to the changes of radiosonde types and thus can be of natural origin. Besides, the latter of these breaks does not concern temperature series, the parameter of major importance for the present study. Finally, according to Zaitseva (1991), systems "Malakhit-A-22" and "Meteorit-PK3-MAP3" do not produce noticeable systematic errors for main meteorological parameters in the troposphere (except for air humidity), which also suggest the absence of methodological inhomogeneity in these balloon-sounding series for the period of 1962–1992.

As for air humidity, the radiosonde measurements in Antarctica cannot be regarded as informative due to the poor work of humidity sensors at low temperatures [Burova et al., 1990; Zaitseva, 1991; Schverdtfeger, 1987]. It is thought that errors of relative humidity values are as large as $\pm 50\%$ [Burova et al., 1990]. In particular, comparison of balloon-sounding humidity data from South Pole with value of total water content as obtained from atmosphere opacity in sub-millimeter range showed that aerological data underestimate water vapor content roughly by ratio of two [Chamberlin, 2001].

II.1.3. Snow accumulation rate.

The most reliable data on snow accumulation rate in the vicinity of Vostok Station are available due to observations on a snow-stake network set up by N.I. Barkov in January 1970 about 1.5 km to the north from the station (Fig. 10). The stake network represents two perpendicular profiles (NS and WE), each being 1 km long. The total amount of stakes is 79, the distance between adjacent stakes is 25 m and one stake (N 60) is common for the both profiles. Each month (since 1996 only once a year) snow build-up and density are measured near the stakes. In spite of lacks of observations in some months due to hard weather conditions, the data series obtained on the Vostok stake network are unique because of their period of observations and representativeness. Annual accumulation at the stake network is calculated as the product of the network-mean snow build-up for the period from December of preceding year to December of current year (corrected for snow thickness settling) and the network-mean snow density of the upper 20-cm layer of snow measured in December of the current year [Barkov and Lipenkov, 1978, 1996; Ekaykin et al., 1998]. The largest methodological problem in using the data from the stake network is the fact that the network is intersected by the route of the annual tractor traverses Vostok–Mirny (Fig. 10). It was shown that to eliminate the inhomogeneity of the snow accumulation series it is sufficient to reject data from the two stakes closest to the route (from down-wind side), N 68 and N 69 [Ekaykin et al., 1998].

Also, by the beginning of this study a number of snow accumulation data were available, as a result of observations at the meteorological site, as well as from stratigraphic and geochemical studies in pits (including 10-m pit *vk10*) and shallow boreholes. Part of these data was published [Averianov, 1969a, 1978; Vilensky and Emelianov, 1971; Vilensky et al., 1970; Vinogradov and Lorius, 1972; Dansgaard et al., 1977].

During the field works carried out by the author together with his Russian colleagues in the frame of present studies (see below), the accumulation data from two 3-m pits (*st61* and *st73*) and two 12-m pits (*vk99* and *st30*) were obtained. Snow accumulation series in four other pits were constructed according to the studies carried out by Russian and French glaciologists (V.Ya. Lipenkov, J.R. Petit, J.M. Barnola and others) in 1984 (*vk14*) and 1998 (*vk31*, *vk33* and *vk34*). Locations of the pits are shown in Fig. 10. Methods of stratigraphic investigations are described in Sections II.2.1 and II.2.2.

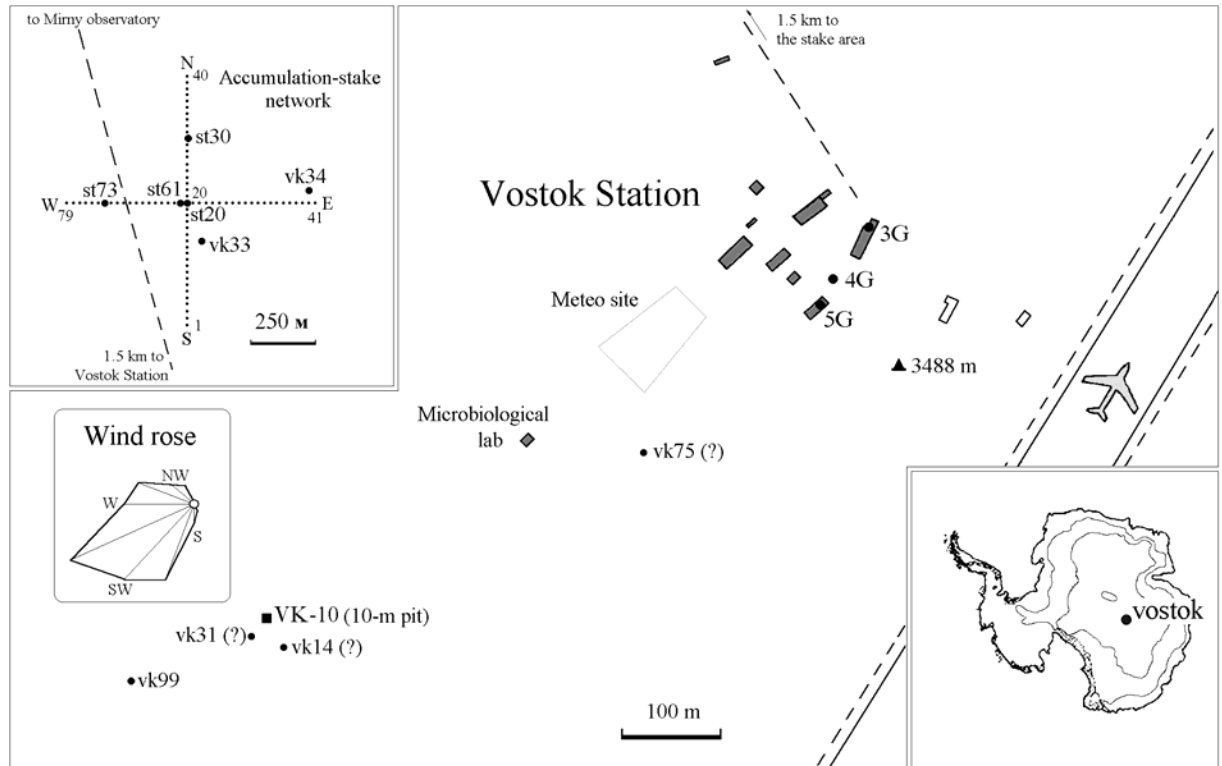


Fig. 10. Scheme of Vostok Station's vicinities. The circles denote the location of shallow and deep pits. The question marks point out pits with doubtful locations (from [Ekaykin et al., 2002]).

II.1.4. Isotope composition of snow.

As pointed in Section I.2.2, by the beginning of the present study only three papers have been published on isotope composition of snow precipitation [Gordienko et al., 1976] and snow cover [Nikolaev and Kolokolov, 1993; Dansgaard et al., 1977] in the vicinity of Vostok Station.

During the present study isotope data were obtained from 6 shallow and 2 deep pits (samples from 4 pits were taken by the author in 1998–2001, and from the other 4 pits by Russian and French glaciologists in 1984 and 1998). See Sections II.2.2 for the methods of snow sampling and II.3.1 for the methods of laboratory measurements of isotope composition.

In January 2000 near each stake of the stake network two snow samples (with depth of 10 and 34 cm) were taken with the aim to study of the spatial distribution of snow isotope composition.

This dataset of snow isotope composition was also completed by 120 samples of precipitating and blowing snow collected during the period from December 1999 to December

2000 by the author and by meteorologist of Vostok V.A. Persky (specially for this study) in order to study the modern seasonal variability of precipitation isotope composition.

II.2. Field works

Field works described below have been carried out with the author's participation during summer seasons of 44th, 45th and 47th Russian Antarctic Expedition at Vostok (1998/99, 1999/2000 and 2001/02).

II.2.1. Stratigraphic studies in pits.

All the snow sections discussed in the present work were dated according to the results of stratigraphic observations taking into account the probability of annual layer hiatus in the vicinity of Vostok [Ekaykin et al., 1998]. The author used methods developed by a collaborator of AARI V.Ya. Lipenkov (to be published elsewhere) based on stratigraphic investigations of snow thickness in different regions of Antarctica carried out by himself and other glaciologists (among others: Alley (1988), Koerner (1971), Kotlyakov (1961), Shumsky (1955)). The obtained initial datings were then corrected using data on the depths of the following reference layers:

- layers of high beta-radioactivity of 1955 and 1965 [Picciotto and Wilgain, 1963; Crozaz, 1969] (for all the eight pits);
- layer of high conductivity of 1816 containing products of Tambora volcano eruption [Legrand et al., 1987] (for the pits *vk10*, *vk99* and *st30*).

To correct the dating of the upper part of snow thickness in pits *st61*, *st73* and *st30* the results of snow build-up measurements (1970–1999) were also used from stakes N 61, 73 and 30 located about 30 cm from the work wall of the corresponding pits. The data used for snow dating are shown at Figures 11 and 12. To correct the dating below Tambora level, where no reference horizons are available, the probability of annual layer hiatus was used as calculated for each pit for the layer between snow surface and Tambora level.

To reconstruct annual values of snow accumulation based on annual layer thickness data, a stacked experimental snow density profile was used (Fig. 13). Interval 0–1 m of the profile includes data from 17 pits and mini-pits dug in 1980–1997. Interval 1–3 m comprises data from 6 pits (*vk10*, *vk22*, *st61*, *st73*, *vk99* and *st30*), while below 3 m data from only two pits are available (*vk10* for interval 3–10 m and *vk99* for 10–12 m).

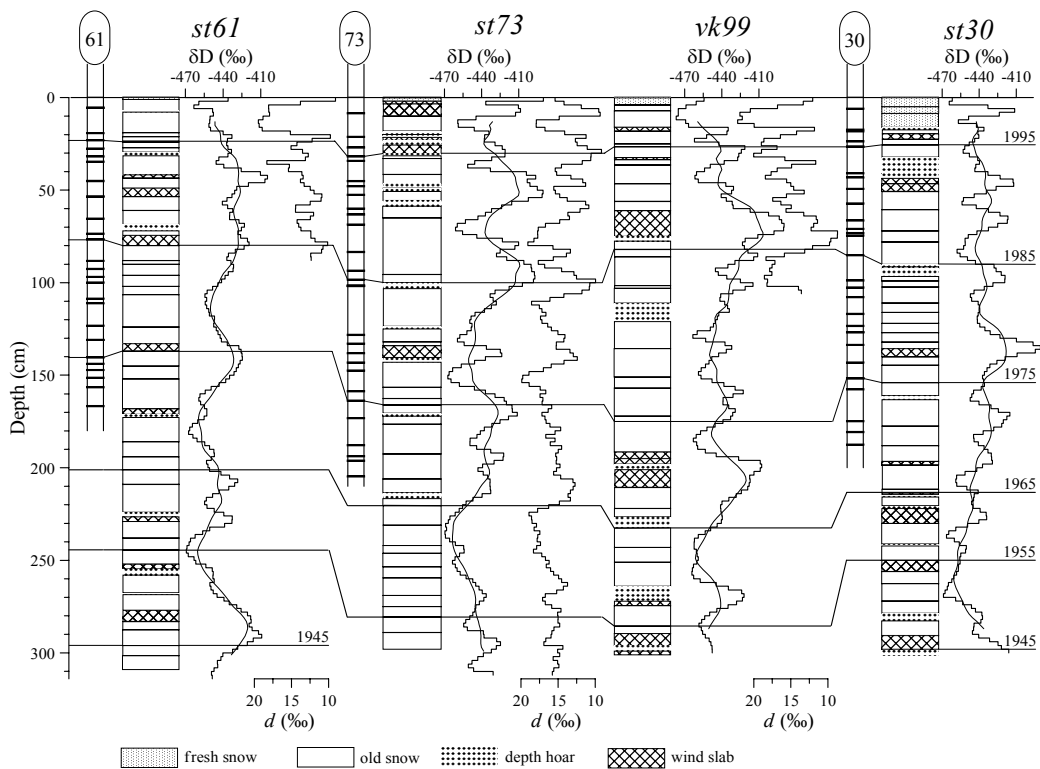


Fig. 11. The results of snow studies (stratigraphy, δD and d) in pits *st61*, *st73*, *vk99* and *st30*. The annual snow build-up values measured at stakes 61, 73 and 30 were corrected for snow settling to correspond with annual layer thickness. The horizontal lines represent isochrones as indicated in the figure. The δD profiles smoothed with a period of 25 cm are shown by thin curves.

The stacked profile of snow density can be approximated by linear equation $\rho = 7.46 \cdot 10^{-5} H + 0.35$, where ρ (g cm^3) is density at depth H (cm).

II.2.2. Snow sampling in pits.

All pit samples for isotope analysis were taken continuously with a depth resolution of 2–10 cm. Methods of sampling, transporting and storing of the samples were chosen to ensure preservation of their isotope composition until the time of laboratory analysis. On the field, samples were put into plastic bags, the edge of the bags were folded several times and sewed with a stapler. Just after their transport to the station, the bags were sealed up and put into a cold storage room (-30°C). To transport the samples to an isotope laboratory, refrigerators were used with temperatures of $-15 - -20^\circ\text{C}$. For further storing, samples were melted in the sealed bags, poured into special hermetic containers and then refrozen. Just before the analysis samples were melted in these containers and then water samples were introduced into the mass-spectrometer.

The same methods were used for all the other isotope samples of precipitating, blowing or deposited snow.

Beta-radioactivity samples were taken continuously with resolution of 5 cm in intervals 1.5–3 m of each pit with the aim of absolute dating of snow deposits (see Section II.1.3). For sampling and transportation of the samples the same methods were used as for the isotope samples. The samples were stored until the measurements in the sealed plastic bags without melting. The weight of each sample was not less than 200 g.

Samples for liquid conductivity of snow were taken continuously with a resolution of 3 cm in the depth intervals 9–11 m of pits *vk99* and *st30*. To avoid samples contamination, special precautions were taken during pit digging (for example, nobody was allowed to walk upwind from the pit). Just before sampling the surface of snow wall was cleaned with a shovel. For sampling itself, the author put on special sterile gloves. Samples were put into special sterile chemical containers which had been opened just before taking the probe, then immediately closed and opened again only in the laboratory just before the measurement. The samples were transported and stored under cold temperature (-15 – -20°C) in the same containers without melting.

The profiles of isotope composition, total β -activity and liquid conductivity of snow from pits *st61*, *st73*, *vk99* and *st30* are shown at Figures 11 and 12.

II.2.3. Sampling of precipitating and blowing snow.

For sampling of precipitating and blowing snow two snow traps were set up about 30 m upwind from the station: one was at the level of snow surface (to collect blowing snow), the other was about 1.5 m above the surface (for precipitation). Snow samples were collected after each snowfall. Samples were collected into special hermetic plastic containers which were then kept in cold conditions. Transportation and storage of the samples was the same as for the isotope samples from pits (see above).

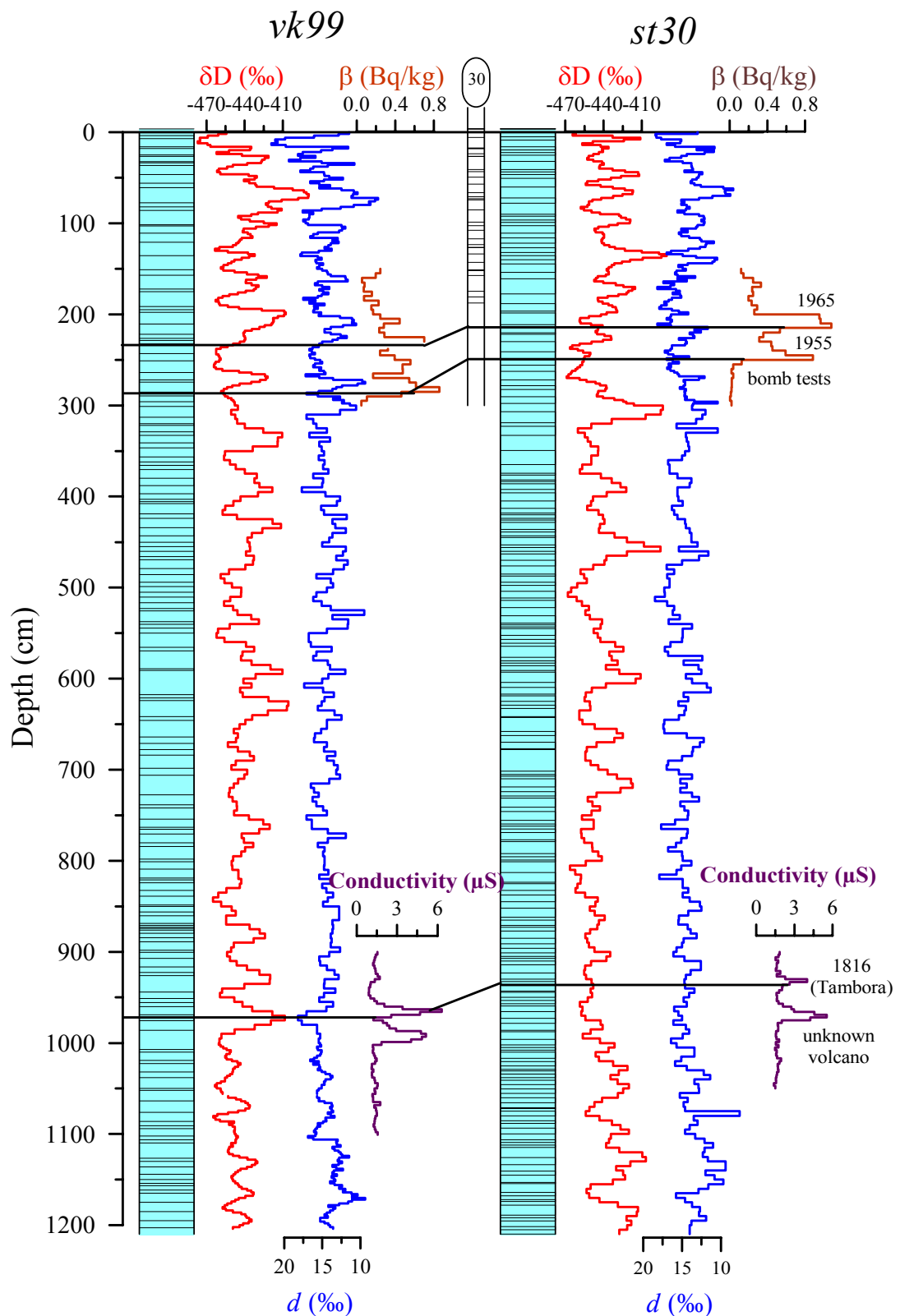


Fig. 12. The results of snow thickness studies (layer boundaries, isotope composition (δD and d), total beta-radioactivity and snow liquid conductivity) in pits *vk99* and *st30*. Values of annual snow build-up at stake 30 are corrected to take into account snow settling. The horizontal lines represent reference horizons of 1955, 1965 and 1816 (from [Ekaykin et al., 2003] with modifications).

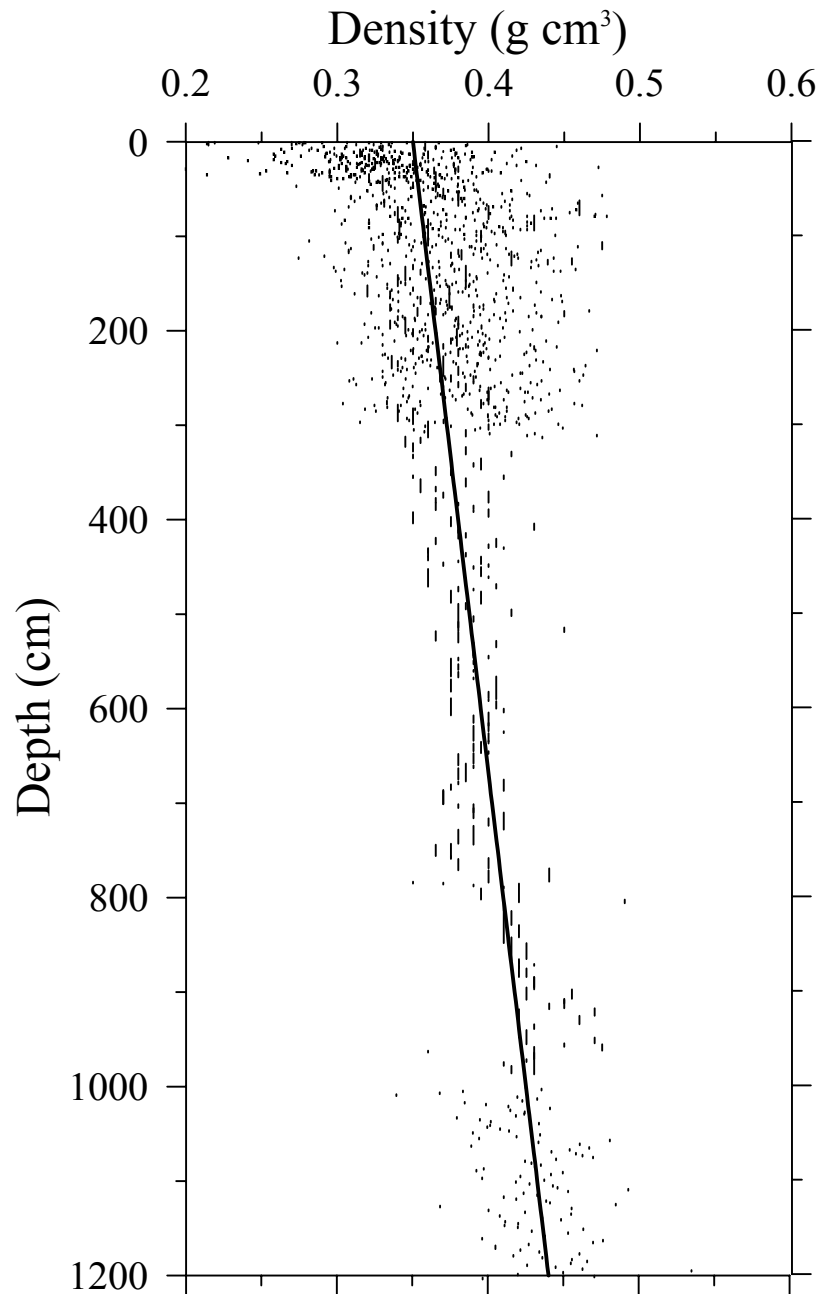


Fig. 13. Stacked profile of snow density at Vostok Station. Approximating line is obtained by least squares method (from [Ekaykin et al., 2003] with modifications).

II.2.4. Construction of new snow accumulation-stake network.

In December 1998 a new stake network was set up to the west of the old one. The observations at the both networks were made at the same date and using the same methods. The design and size of the new stake network is identical to those of the old network, and their WE profiles represent single profile with the total length of 2 km. The only difference between the two constructions is that the stakes of the old network are made of metal, while those of the new one are made of bamboo. In the present work the data from the new stake network are not used

because some methodological problems related to compatibility of the data from the both network have to be solved.

II.2.5. Snow surface leveling.

In December 1999 we carried out a leveling of the snow surface near each stake of the old and the new stake networks in order to reveal relationship between snow relief and spatial variability of snow accumulation rate and isotope composition. Before the survey both systematic and random errors of the observations were determined empirically which were proved to be satisfactory for the purposes of this study.

II.3. Laboratory measurements

II.3.1. Isotope measurements.

All the isotope measurements made in the frame of the present work have been carried out in Laboratoire des Sciences du Climat et de l'Environnement (LSCE), Saclay, France (deuterium and oxygen 18) and in Department of Geophysics of Niels Bohr Institute of Copenhagen University, Denmark (oxygen 18) with the author's participation.

The principle of mass-spectrometry is related to separation of moving ions by their masses as a result of magnetic and electrical fields application. The differences in measurements of isotope composition of various elements mostly concern procedures of probe preparation before its introduction into the mass-spectrometer.

In particular, to determine the concentration of oxygen 18 in water samples the probe to be measured is isotopically equilibrated with carbon dioxide under constant temperature. The isotope composition of oxygen in the used CO₂ is very well known. The CO₂ then passes to mass-spectrometer and after the measurement its isotope composition is recalculated into the isotope composition of initial water sample. This method is currently used in both LSCE and Niels Bohr Institute. To determine the concentration of deuterium, the analyzed water sample is evaporated and water vapor molecules are decomposed by reaction with hot uranium (this method is used at LSCE), chromium or zinc to release molecular hydrogen which then passes to mass-spectrometer. Due to a number of reasons, the latter method is technically much more complicated. In the most of modern mass-spectrometers both preparative and analytical parts are coupled one to another, and the processes of probe injection into mass-spectrometer and analysis are automatized.

The analytical part of mass-spectrometer consists of three sections: 1) an ion source where molecules of studied elements are turned into ions under a beam of electrons; 2) mass-analyzer where ions are accelerated in an electrical field and then separated into different beams according to their masses in a magnetic field; 3) collectors of ions. The voltage at the different collectors depends on the amount of ions of different masses and thus on the initial concentration of the different isotopes in the samples. To obtain the final result of the isotope composition several corrections are applied to the data produced by the mass-spectrometer. These corrections take into account the reproducibility of apparatus, its "memory" and also the concentration of parasitic ions received by the collectors. To increase the reliability of the results, several "work standards" (laboratory water with precisely known isotopic composition which does not much differ from that of the measured samples) are put into the measured series together with the samples. The isotope composition of the samples is calculated relatively to that of the "work standard" and then re-calculated in the conventional δ notation (i.e., relatively to the international SMOW standard). Detailed description of the mass-spectrometer principles can be found in [Delmotte, 1997; Stievenard et al., 1994; Vasil'chuk and Kotlyakov, 2000; Vimeux, 1999].

The modern mass-spectrometers allow determination of isotope composition of water samples with precision of 0.5 % for deuterium and 0.05 % for oxygen 18. The reproducibility (random error) of measurements carried out during the present study was in most cases within 0.6–0.8 % for δD and 0.04–0.06 % for $\delta^{18}O$, which is quite sufficient for the purposes of the work. This gives the precision of deuterium excess values between ± 0.7 and ± 0.9 ‰, that is, one order of magnitude less than the natural variability of this parameter (see Figures 11 and 12). The worst accuracy was obtained for δD values for the precipitation samples collected in 2000: ± 1.5 ‰, accuracy for d values being ± 1.6 ‰. The reason for this was comparatively poor state of the mass-spectrometer during the measurements. Though, this value is still very small comparing to the absolute magnitude of seasonal changes of precipitation isotope composition at Vostok (165 ‰). According to our estimations, the systematic error of the measured isotope value does not exceed ± 1.3 ‰ for δD , ± 0.11 ‰ for $\delta^{18}O$ and ± 1.2 ‰ for deuterium excess (see Annex C for the details).

II.3.2. Measurements of beta-radioactivity.

Total β -radioactivity of the snow samples was measured in Laboratoire de Glaciologie et Geophysique de l'Environnement (LGGE), Grenoble, France, with the participation of the author. Sample preparation for measurements consists of snow melting and pumping of the water during at least 2 hours through ionic filters where all the impurities are deposited. Then the filters

are dried and put into the β -particles counter placed into an underground room to protect it from background radiation. One of the counter's collectors is left empty to account for the background. Then, beta-particles are counted during 2–6 hours which allows determining total beta-radioactivity (in Bq kg⁻¹). The technique is described in [Delmas, Pourchet, 1977; Pinglot, Pourchet, 1979; Pourchet et al., 2003].

II.3.3. Measurements of liquid conductivity.

Measurements of liquid conductivity of snow were carried out in LGGE with the author's participation. In order to avoid samples contamination, the measurements were made in a chemical clean room using special clothes. Before the measurements the samples were melted and warmed up to the room temperature. The conductivity was determined by conductometer CD78 produced by TACUSSEL[®].

II.4. Conclusion of Chapter II

In this chapter the review is given of the meteorological and balloon-sounding data, as well as data on isotope composition and accumulation of snow, available prior to the beginning of the present study and obtained during its implementation. It was revealed that balloon-sounding series contain methodological inhomogeneity during the period 1958–1961. Methods of field works (stratigraphic and geochemical investigations in snow pits, sampling of precipitating, blowing and deposited snow) and laboratory analyses (isotope studies, measurements of total beta-radioactivity and liquid conductivity of snow) are described. In general, it can be concluded that the dataset collected by the author contains all the information needed for completing the present study:

- 1) Meteorological data (mean 10-day, monthly and annual values) for the period 1958–2002;
- 2) Balloon-sounding data (daily, mean 10-day, monthly and annual values) for the period 1963–1991;
- 3) Data on snow accumulation rate from stake network (1970–2001), from stratigraphic and geochemical studies in a number of pits and shallow boreholes (for the last 200 years);
- 4) Data on snow isotope composition of snow thickness from 8 pits (for the last 60 years from 6 pits and for the last 200 years from 2 pits), precipitating and blowing snow (December 1999–December 2000), as well as data on spatial distribution of isotope composition of surface snow;

5) Results of snow surface leveling in the area of the stake network.

Most of these data have been obtained by the author during the present study.

III. METEOROLOGICAL CONDITIONS OF SNOW FORMATION

This chapter addresses the following questions:

- 1) How much do different precipitation types contribute to the total amount of precipitation at Vostok?
- 2) What is the difference between precipitation and accumulation rates?
- 3) Where effective level of precipitation formation is situated and what are the meteorological conditions at this level?
- 4) How large is the difference between mean annual values of the meteorological characteristics and their mean values weighted by the amount of precipitation?

III.1. Contribution of different precipitation types in total precipitation amount

As pointed out in Section I.1.3, in central Antarctica precipitation are represented by three types: precipitation from clouds, precipitation from clear sky (diamond dust) and rime. As for the latter, it is obvious that condensation (inverse sublimation) can exceed sublimation only if snow surface is colder than near-surface air, that is, in winter. On the other hand, during the cold season water content of air is very low (Fig. 5), which does not suggest large amount of rime in this area. This conclusion is confirmed both by direct measurements and by calculations based on turbulent diffusion method. In the first case total sum of rime during the winter season was 0.06 g cm^{-2} [Averianov, 1990] while in the latter case it was 0.018 g cm^{-2} [Artemiev, 1976]. These figures represent, correspondingly, 3 and 1% of mean annual sum of snow accumulation. Thus, contribution of this precipitation type at Vostok is negligible. On the other hand, measurements fulfilled at Japanese Dome Fuji Station (meteorological conditions similar to those of Vostok) showed that during period from March to October monthly sums of condensation was $0.03\text{--}0.1 \text{ g cm}^{-2}$ [Motoyama, personal communication, 2003], that is, about an order of magnitude larger than at Vostok. It is not clear whether contribution of diamond dust is excluded from these values or not.

Review of available literature (see Section I.1.3) suggests that relative amount of cloud precipitation and diamond dust in central Antarctica has not still been determined with sufficient accuracy. Contribution of the latter type is estimated from 30 to 90% of total precipitation sum [Averianov, 1990; Radok, Lile, 1977; Schwerdtfeger, 1987]. Thus, we made an attempt to determine amount of both these types using data on monthly accumulation at the stake network of Vostok Station (1970–1995).

First of all, we determined the relationship between the precipitation amount and the surface mass balance (or net accumulation) at Vostok Station. In central Antarctica, where no summer melting occurs, the snow accumulation rate (a), obtained from the snow measurements at the stake network, is equal to:

$$a = p - s + q,$$

where p is the precipitation rate (including rime), s is the snow sublimation and q is net snow redistribution by wind (amount of snow blown into the studied area minus snow blown away from this area). Generally, one cannot assume the q value to be zero, simply because the horizontal mass flux (blowing snow) is, at least, 10^5 times more intense than the vertical flux (precipitation, condensation and sublimation) [Averianov, 1990], so very slight spatial variations of the blowing snow intensity can significantly alter the local mass balance. Indeed, it is widely recognized that vast areas of the Antarctic continent (so-called "katabatic wind belt" in the lower part of Antarctic ice sheet slope) are subjected to intensive wind erosion, so the q parameter is strongly negative there [e.g., Gallee et al., in preparation; Lipenkov et al., 1998]. However, for the interior parts of Antarctica with flat and relatively smooth surfaces and light wind regime it is believed that the snow redistribution by wind does not change the long-term average accumulation rate on the sufficiently large areas (of the order of 10^1 – 10^2 km²) [Averianov, 1990]. At a first approximation, the Vostok vicinities meet these conditions (though, as will be shown in Chapter VI, we still cannot totally exclude the influence of large relief forms on the snow accumulation in this area), so we simply assume that snow accumulation rate here is equal to the precipitation rate minus sublimation rate:

$$a = p - s.$$

Thus, snow accumulation values for the "warm" period of year (November–February) were corrected for the snow sublimation to produce the precipitation values.

As an estimation of sublimation the average of calculated values by turbulent diffusion method [Artemiev, 1976] and direct measurements of V.K. Nozdryukhin [Averianov, 1990] were taken. The calculations based on turbulent diffusion give values for November, December and January (for 1966 and 1969) equal to 0.38, 1.9 and 1.9 mm, correspondingly (data for February is absent). At the same time, according to the direct measurements of D.N. Dmitriev, sublimation for the same months of 1982 was one order of magnitude higher: 5.25, 14.0 and 5.6 mm [Averianov, 1990] which represents in total 110 % of the annual snow accumulation in this region [Barkov, Lipenkov, 1996]. However, these data are not perfect because frozen water was used as sublimating sample, which is not representative in terms of natural conditions. On the contrary, data of V.K. Nozdryukhin (summer 1961/62), who used for his measurements actual

samples of surface snow, are much more consistent with the theoretical ones: 0.36, 0.50 and 0.56 for November, December and January, correspondingly) [Averianov, 1990]. Value of sublimation in February was taken equal to that in November assuming symmetric inter-annual distribution of sublimation following symmetric distribution of air temperature and radiation balance of the surface. This assumption is excusable taking into account uncertainty of our knowledge about the components of mass-balance in central Antarctica.

It was then suggested that sublimation during a given month of year is constant over the whole period of interest. Mean monthly values of snow accumulation and sublimation and resultant calculated values of precipitation are listed in Table 4. We should note that sublimation values shown in column 3 of Table 4 are within the same order of magnitude to the corresponding values measured at Dome Fuji Station: 0.16–0.20 g cm⁻² for December and 0.19–0.20 g cm⁻² for January, which gives an indirect support to our estimations [Motoyama, personal communication, 2003].

Table 4. Mean (for the period 1970–1995) monthly and annual values of snow accumulation and sublimation, and resultant calculated values of precipitation, mm w.e.

Month	Accumulation	Sublimation	Total Precipitation	Precip. from clouds	Ice crystals
<i>I</i>	<i>2</i>	<i>3</i>	<i>4</i>	<i>5</i>	<i>6</i>
I	1.2±2.0	1.2	2.4	1.5	0.9±1.5
II	1.4±1.7	0.4	1.8	0.1	1.7±2.6
III	2.0±1.5	0	2.0	0.4	1.5±1.8
IV	2.3±2.0	0	2.3	≥0	≤2.3±2.1
V	1.9±1.5	0	1.9	1.0	0.9±0.6
VI	1.7±1.4	0	1.7	0.1	1.6±1.8
VII	2.8±1.6	0	2.8	0.6	2.2±0.9
VIII	2.1±1.5	0	2.1	0.4	1.7±1.4
IX	3.0±1.9	0	3.0	0.6	2.4±1.4
X	2.8±2.1	0	2.8	0.9	1.9±1.7
XI	0.7±1.7	0.4	1.1	≥0	≤1.1±1.4
XII	0.2±1.7	1.2	1.4	1.0	0.4±1.8
Year	22.9±4.4	3.2	26.1	6.9	19.2±5.8

Error intervals refer to the inter-annual variability (1σ) of the corresponding parameters.

At the next step the data on the amount of days with precipitation from clouds were involved. It was suggested that during the months when no precipitation from clouds (S) were observed, all the precipitated snow is formed by diamond dust (DD); alternatively, during the months with precipitation from clouds, monthly sum of precipitation is equal to $S + DD$. Thus, we choose the months when no precipitation from clouds were observed and for each of these months precipitation sum was calculated equal to accumulation of snow at the stake network plus sublimation of snow. Then for all the 12 months of year mean multi-year values of DD were calculated (column 6 of Table 4). Then, mean multi-year monthly values of precipitation from clouds (column 5) were determined as difference between mean multi-year values of total precipitation (column 4) and mean multi-year values of precipitation from clear sky (column 6). The obtained results are presented at Figure 14.

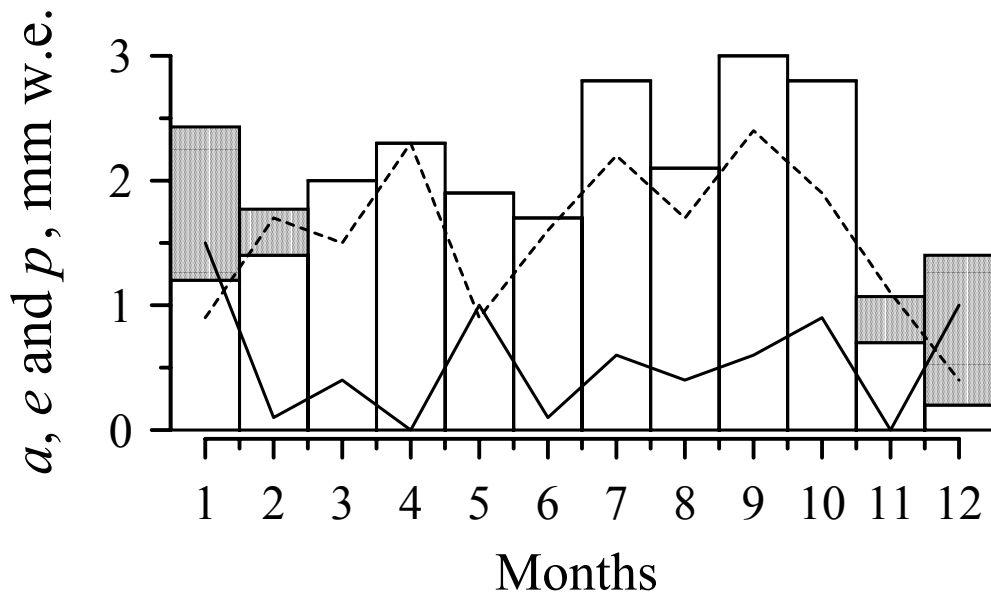


Fig. 14. Seasonal variations of total snow precipitation (bars) including precipitation from clear sky (dashed line) and clouds (solid line). Shaded parts of the bars represent the portion of precipitation which is sublimated during the summer season. See text for the details.

The seasonal distribution of precipitation appears to be much more homogeneous than that of snow accumulation because the summer minimum of the latter is mainly due to snow sublimation during the warm period. The most important conclusion for our study is that $\frac{3}{4}$ of total precipitation amount (between 61 and 89 % with the confidence level of 95 % according to the error bars given in the last row of Table 4) are formed by diamond dust and only $\frac{1}{4}$ is due to cloud precipitation. Keeping in mind the large relative errors of the monthly precipitation values, we avoid discussing the seasonal variability of the mass-balance components. Still, it is worth

noting the minimum clear sky precipitation in summer (December and January) probably due to weaker inversion and decreased subsidence in the atmosphere above Antarctica, as well as due lower possible water content in colder air (see Section I.1).

Such a small contribution of cloud precipitation could be explained by low occurrence of the corresponding events (on average, 37 days with cloud precipitation per year according to the meteorological data over the period 1963–2001), and also by the low intensity of these precipitation which even leads to difficulties in distinguishing between cloud precipitation and diamond dust [Averianov, 1990; Schwerdtfeger, 1987].

III.2. Temperature of condensation

It was in 1977 that English scientist G. Robin in his work "Ice cores and climatic changes" first suggested that condensation temperature T_C in Antarctica is within 4°C equal to the temperature T_i at the top of surface inversion layer [Robin, 1977]. The idea behind this assumption is that the inversion layer contains comparatively large amounts of water vapor, so the air in this layer becomes supersaturated in respect to ice as soon as radiation cooling starts leading to condensation and precipitation formation. Approximating T_C by T_i has been widely used [Ciais and Jouzel, 1995; Jouzel and Merlivat, 1984; Petit et al., 1999 and many others] in isotope studies during the last 25 years, although no experimental evidence for the equality of these parameters was available. Let us note incidentally that in the coastal areas the level of condensation (1–2 km above sea level) is well above the upper boundary of thin surface inversion layer, which makes doubtful using spatial dT_i/dT_S ratio instead of dT_C/dT_S ratio (see the Section I.3.4). Below an attempt is made to determine the condensation temperature for Vostok conditions, using the available dataset of balloon-sounding data, and to validate the common practice of approximating T_C by T_i (for the conditions of Vostok).

In case of precipitation from clear sky the level of condensation (more correctly, layer of condensation) is likely correspondent to the quasi-isothermal layer just above the layer of surface inversion (Section I.1.3). Thus, the best estimation of T_C for diamond dust is the mean temperature of this isothermal layer, which is generally in agreement with the Robin's assumption. According to the balloon-sounding data, mean temperature of this layer is $-39\pm 1^\circ\text{C}$ (where $\pm 1^\circ\text{C}$ is the amplitude of the vertical changes of the temperature in the isothermal layer) (Figure 15).

As for the precipitation from clouds, condensation temperature is in first approximation equal to the temperature at the lower boundary of clouds, data on which are available from

balloon-sounding observations. Comparison of seasonal changes of height of lower cloud boundary (and its temperature) with height and temperature of the upper boundary of surface inversion is given in Table 5 and at Figure 16.

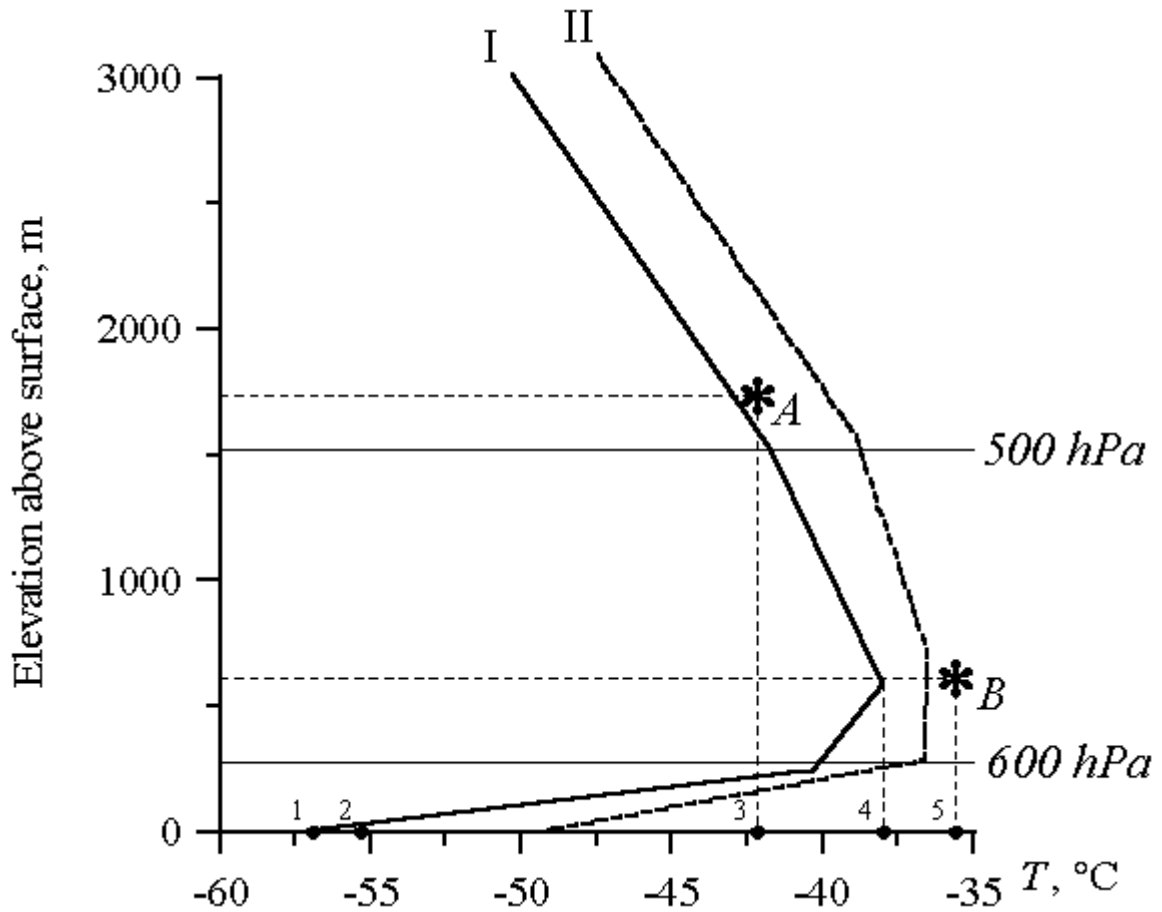


Figure 15. Mean annual (1963–1991) vertical profile of the air temperature (curve I) and mean temperature profile during the cloud precipitation events (curve II). *A* and *B* are mean height and temperature of middle and lower clouds, correspondingly. 1 and 4 – temperatures at the bottom and the top (T_i) of the inversion layer, correspondingly; 2 – near-surface air temperature (T_s); 3 and 5 – mean temperatures at the lower boundary of middle (A_s , A_c) and lower (S_t , S_c) clouds, correspondingly. Also shown is the mean height of 500 hPa and 600 hPa isobaric surfaces.

Figure 16 and Table 5 suggest that the height of lower boundary of the lower clouds (S_t and S_c) (4100 m) is within its uncertainty equal to the height of upper boundary of surface inversion (4060 m during the days with clouds and 4080 m for all the days). At the same time, temperature at the lower boundary of lower clouds is systematically lower (on average by 2.1°C) than T_i during the days with clouds, but systematically higher (by 2°C) than mean annual value

of T_i . This implies that when the lower clouds are observed, the temperature in the warmest atmospheric layer is higher than its mean annual value by about 4°C, which is likely related to warm advection. In general, one can note that condensation temperature of precipitation falling from lower clouds is equal (within its uncertainty) to the temperature at the upper inversion boundary, the difference between these two parameters being slightly less in summer than in winter.

Table 5. Mean (for the period 1963–1991) monthly values of height of lower cloud boundary (and its temperature), as well as height and temperature of the upper boundary of surface inversion

Month	Lower clouds				Middle clouds				Upper inversion boundary	
	H_C, m	$T_C, ^\circ\text{C}$	H_i^*, m	$T_i^*, ^\circ\text{C}$	H_C, m	$T_C, ^\circ\text{C}$	H_i^*, m	$T_i^*, ^\circ\text{C}$	H_i, m	$T_i, ^\circ\text{C}$
I	3920	-30.8	3940	-28.8	5050	-32.9	3860	-28.4	3903	-30.1
II	4125	-36.8	4040	-32.6	5750	-	4090	-34.7	4106	-33.9
III	4040	-36.0	3855	-34.0	5580	-	4140	-37.4	4154	-38.0
IV	4090	-38.8	4010	-36.5	4650	-38.3	4030	-34.3	4149	-40.2
V	3830	-33.4	4145	-30.5	5330	-38.1	4200	-34.3	4136	-40.8
VI	4160	-33.7	4440	-33.8	5580	-46.6	4090	-35.2	4147	-40.7
VII	4300	-44.0	4025	-42	4940	-	4160	-35.1	4142	-41.9
VIII	4300	-	4340	-33.6	5370	-49.8	4010	-40.3	4133	-42.4
IX	4110	-36.5	4240	-35.6	5060	-42.8	4060	-38.3	4122	-42.5
X	4125	-	3915	-37.6	4960	-48.0	3985	-41.1	4102	-40.1
XI	4080	-	3770	-35	5340	-46.0	3935	-38.2	4028	-35.9
XII	4110	-30.5	4000	-27.5	5130	-37.0	3990	-30.0	3843	-30.4
Year	4099	-35.6	4060	-34.0	5228	-42.2	4046	-35.6	4080	-38.1

H_C and T_C are the height of lower cloud boundary and its temperature; H_i^* and T_i^* are the height of upper inversion boundary and its temperature during the days with corresponding clouds; H_i and T_i are the mean values of height of upper inversion boundary and its temperature.

As for the middle clouds, their lower boundary height is considerably (more than by 1 km) larger than upper limit of surface inversion and is on average equal to 5200 m not revealing considerable seasonal changes (similar to lower clouds). Temperature at lower boundary of middle clouds is by more than 6 °C lower than at the top of inversion during the day with clouds

and by 4 °C lower than mean annual T_i value. It should be noted that real difference between condensation temperature in clouds and T_i can be even larger if taking into account that precipitation is formed in the whole cloud layer rather than at its lower boundary. The points corresponding to the height and temperature of lower boundary of lower and middle clouds are shown at Figure 15 by characters *A* and *B*.

On average, according to meteorological data, occurrence of precipitation from lower and middle clouds is nearly equal. But, taking into account exponential relationship between moisture content in air and temperature [Dinamicheskaya meteorologiya, 1976], one can suppose that mean weighted cloud condensation temperature is closer to the temperature at the lower boundary of lower clouds (that is, between points 3 and 5 at Figure 14, but closer to the latter).

In general, according to Figure 15, possible range of mean annual condensation temperature in clouds at Vostok is between -42 and -36 °C, i.e., -39 ± 3 °C (we also assume that inter-annual variability of this parameter falls within this uncertainty). Averaging this value with the above given condensation temperature for the precipitation from clear sky (-39 ± 1 °C) gives the most probable value of the overall condensation temperature at Vostok equal to -39 ± 1.6 °C (calculated taking into account that most of precipitation is formed in the inversion layer (diamond dust)). The latter is correspondent, within its uncertainty, to the mean annual temperature at the inversion top (-38.0 ± 0.6 °C) thus confirming Robin's assumption (see the beginning of this section) for the Vostok's conditions.

Curve II at Figure 15 shows the mean vertical air temperature profile during the days with cloud precipitation. As expected (see review in Section I.1), the temperature during precipitation events is higher at any level than its mean annual value. However, this difference is not constant in altitude changing from 6 °C near the surface to the minimum of 1.5 °C at the top of inversion and then increasing again up to 3 °C in the free atmosphere. This implies considerable weakening of inversion due to green-house effect of clouds preventing radiation cooling of the surface.

Interestingly, profile II does not pass through points *A* and *B*. This can be explained by the fact that lower and, especially, middle clouds do not always give precipitation. On the other hand, according to the meteorological data, about 2/3 of all "cloud" precipitation were observed in the presence of only upper clouds (no lower or middle ones). Possibly, in many (or most) of these cases the observer erroneously recognized diamond dust as cloud precipitation. This example underlines once more the complexity of observing precipitation in central Antarctica (already mentioned in Sections I.1.3) and strengthens the need to be very cautious when treating the data on precipitation amount and their occurrence. In our study, we used the data on cloud

precipitation days only for those days when lower or middle clouds were observed; in other cases (upper clouds only) we considered "cloud precipitation" as "diamond dust".

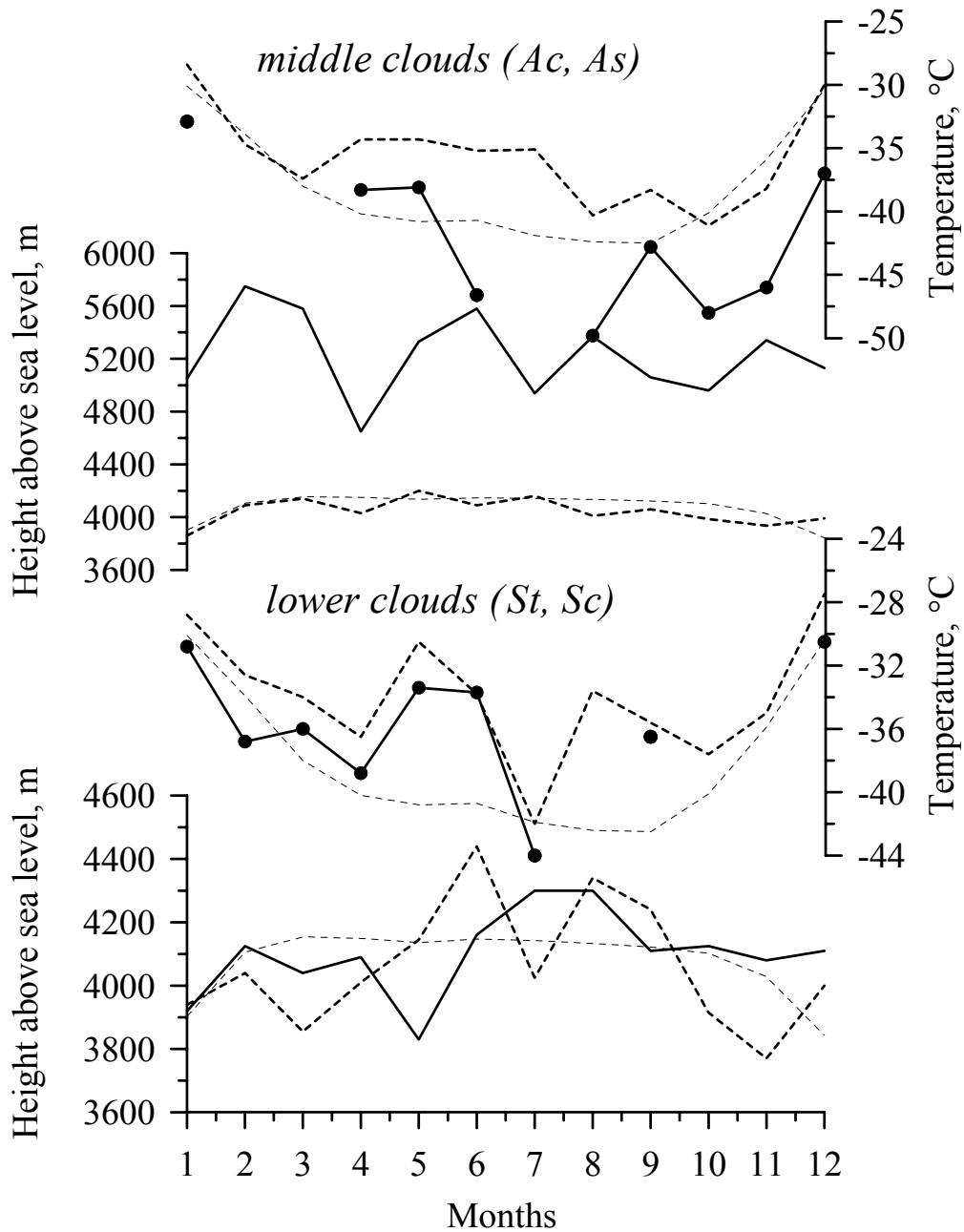


Fig. 16. Comparison of seasonal changes of height and temperature of lower boundary of lower and middle clouds (solid lines) with height and temperature of upper boundary of surface inversion during days with corresponding clouds (dashed lines). Thin dashed lines represent the mean multi-year seasonal variations of height and temperature of the inversion top. The data used for the figure are given in Table 5.

Taking into account that most of precipitation is formed under clear-sky conditions (diamond dust) we argue that precipitation-weighted condensation temperature at Vostok is not

significantly biased by its mean annual value. This is due to the fact that diamond dust is observed practically every day throughout the year. So, further in our study we use mean annual values of temperature as the most reliable approximation of precipitation formation conditions (and thus likely of precipitation isotope composition).

III.3. Conclusion of Chapter III

In this study, a first attempt was made to use monthly data on snow accumulation at the Vostok stake network in order to estimate the contribution of different precipitation types into the overall precipitation amount. In central Antarctica, it is only at Vostok Station that this task could be performed with sufficient accuracy because only here the reliable (obtained on sufficient number of observation points over sufficiently large area and using the same technique) data on snow accumulation are accompanied with meteorological information over comparatively long period of time. The obtained results suggest that most of precipitation, roughly 74 % (between 61 and 89 % with probability 95 %), falls from clear sky as diamond dust, while clouds gives only about 26 %. The contribution of rime (condensation from air directly to the snow surface) is negligibly small.

To estimate the mean annual precipitation rate we assumed that in the area of Vostok the surface mass balance equals to precipitation minus sublimation. Thus, we suppose that the mass balance term related with blowing snow is, on average, zero. The data on the sublimation rate was taken from the available literature. To make the obtained results more reliable, new mass balance observations (including precise measurements of sublimation/condensation and of blowing snow intensity) have to be carried out at Vostok Station during a full annual cycle.

The overall condensation temperature at Vostok Station equals to -39 ± 1.6 °C (where ± 1.6 °C is an estimated uncertainty that accounts for the different contribution of various precipitation types into the total precipitation amount) and, thus, does not significantly differ from the mean air temperature at the top of the surface inversion layer. We suppose that the mean annual temperature is well representative in terms of isotope composition of snow due to dominant role of diamond dust (observed nearly each day throughout a year) in the total amount of precipitation.

The obtained results suggest that the local temperature recorded in the isotope composition of snow deposits at Vostok Station most closely corresponds to the mean annual air temperature at the upper inversion boundary.

IV. SPATIAL VARIABILITY OF SNOW ISOTOPE COMPOSITION: PLAYGROUND OF WIND

This chapter is devoted to the following topics:

- 1) The main types of snow surface relief forms in central Antarctica;
- 2) The influence of snow redistribution due to wind activity on the formation of vertical profiles of isotope composition of snow thickness;
- 3) Separation of "relief-related" and climatic variations in temporal series of accumulation rate and isotope composition of snow.

As mentioned in Section I.3.6, snow redistribution by wind can result in spatial and temporal variability of snow isotope composition. In this chapter we discuss the spatial variability of snow isotope composition (δ) at Vostok Station and its possible influence on temporal δ oscillations in a single point.

IV.1. Mega-dunes and micro-relief

It has been shown [Black and Budd, 1964; Frezzotti et al., 2002; Gow and Rowland, 1965; Whillans, 1975; Van der Veen et al., 1999] that the quasi-periodical spatial variability of snow accumulation with the wavelengths of 2 to 40 km observed all over the Antarctic ice sheet is caused by the undulations of snow surface. Due to their large horizontal sizes these undulations (conventionally called as "mega-dunes" [Frezzotti et al., 2002]) affect the velocity of katabatic wind in the inversion layer, since the wind speed is proportional to the surface slope. In turn, this produces a redistribution of recently deposited and precipitating snow, because the intensity of near-surface snow drifting (blizzards) is wind-speed dependant. The superposition of the general slope and the snow undulations results in maximum accumulation occurring not at the bottom, but at the down-wind slopes of the hollows where the surface inclination is smaller. This makes the undulations shift with time in upslope direction (and thus upwind). This drift of snow dunes causes non-climatic temporal oscillations of snow accumulation in a given point with the periods of these oscillations being related to the horizontal sizes of the above undulations through the velocity of their drift.

However, the above mechanism is only valid for the undulations with the wavelengths of the order of 2 km or more, because the smaller snow dunes cannot affect the mean katabatic wind velocity [Whillans, 1975].

Also very well known are the so-called micro-relief forms (like sastrugi), their horizontal sizes not exceeding first tens of meters (see, for example, [Kotlyakov, 1961]). Since these snow ripples have comparatively large height-to-length ratio (of the order of $n \times 10^{-1} - n \times 10^{-2}$, that is one or two orders of magnitude higher than for mega-dunes), they play a role of snow-barriers and thus redistribute the snow accumulation by influencing the turbulent component of the near-surface wind. The snow is predominantly accumulated behind barriers, which make the micro-relief forms move downwind. Micro-relief is assumed to be the main source of noise in accumulation time-series obtained in single point (so-called "stratigraphic noise" [Fisher et al., 1985]).

IV.2. "Meso-dunes" signature in spatial snow build-up distribution

Spectral analysis of the spatial series of snow accumulation as obtained at Vostok accumulation-stake network revealed (Fig. 17) 3 types of spatial quasi-periodical snow accumulation waves with the typical wavelengths of about 50–60, 100 and 230–330 m (waves I,

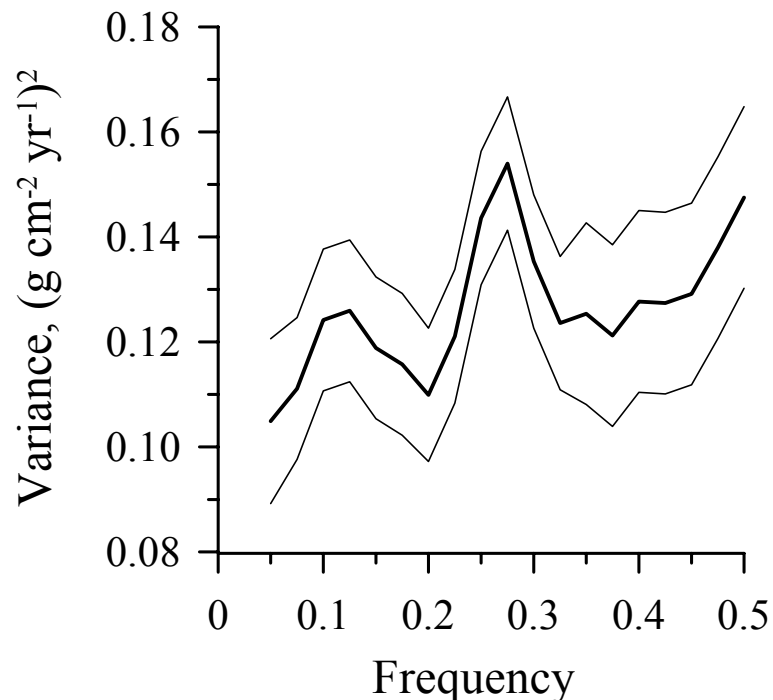


Figure 17. The mean (over the period 1970–1996) spectrogram of the spatial snow accumulation profile measured at the NS section of the Vostok accumulation-stake network. The error bars show inter-annual variability (1σ) of the variance vs frequency function. The frequency is determined as $1/P$, where P is the period measured in distances between the adjacent stakes.

II, and IV, see column 4 of Table 6) [Ekaykin et al., 1998]. Following the above considerations, we associate these waves with not previously described snow surface undulations which horizontal sizes are between those of micro-relief and mega-dunes (let us call them "meso-dunes"). Also, at the series of total snow build-up at the stake area over 30 years (Fig. 18) one can easily see longer accumulation wave with a period of the order of one (or few) kilometer (wave VI, see column 4 of Table 6). Due to its large wavelength, this wave can possibly be linked with a mega-dune undulation.

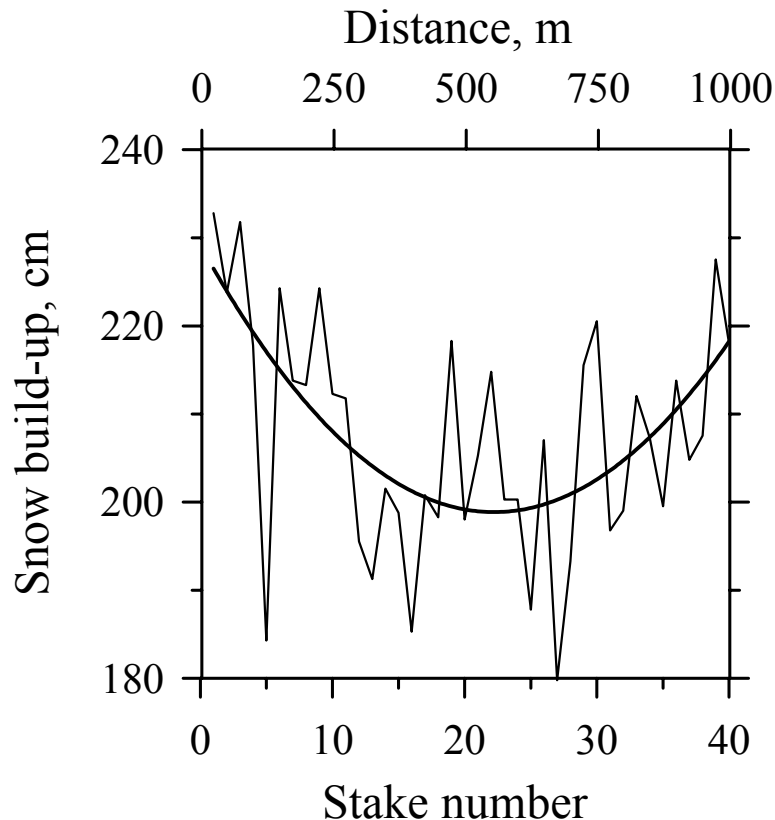


Figure 18. The total snow build-up (cm) at the NS profile of the Vostok accumulation-stake network over the period 1970–2000. The build-up values are corrected for the settling of snow.

The temporal evolution of one type of these waves (wave IV from Table 6) is shown at Figure 19 as an example. One can see the southward (i.e., upwind – see the windrose inserted in Fig. 10) drift of the waves. At present we cannot satisfactorily explain the mechanism of this drift: these waves are too small to affect mean katabatic wind speed and, at the same time, they cannot be snow barriers since their too small height-to-lengths ratio ($n \times 10^{-4}$).

Obviously, as a wave drifts across the stake profile, the snow build-up at a given stake first increases and then, after the wave's crest has passed through the stake, decreases to a

minimum value corresponding to the hollow between two adjacent waves. This forms temporal oscillations of snow build-up at a given stake (the same is valid for a pit or an ice core) not related to climatic variability. We call such variations "relief-related" to distinguish them from purely climatic ones.

Table 6. Periods of oscillations revealed in temporal and spatial series of isotope composition and accumulation rate of snow at Vostok. The typical wavelengths of the snow surface undulations as obtained by the snow surface leveling are also shown (from [Ekaykin et al., 2002] with updates).

Wave	Accumulation			Isotope composition			Relief
	Temporal series, years		Spatial profiles, m	Vertical profiles, cm of snow*	Temporal series, years	Spatial profiles, m	Wave-lengths, m
<i>I</i>	2	3	4	5	6	7	8
I	2.6	2-3	50-60	12-18	2-3	55-75	50-60
II	5.2	4-5	90-110	20-30	4-5	~85	~85
III	10-11	8-11	-	60-80	8-14	-	-
IV	?	~20	230-330	~120	~20	250-330	260-310
V	?	~50	-	~300	~50	-	-
VI	?	>200	≥1000	?	?	≥1000?	≥1000

* mean annual layer thickness is 5.8 cm for the upper 3 m and about 5 cm for the upper 12 m

Based on the drift of accumulation waves with time (as it is shown at Fig. 19), it is possible to calculate the velocity of their movement (e.g., this velocity for the wave IV was found to be about 12 m yr⁻¹) and, thus, to estimate the typical period of relief-related accumulation oscillations in a single point caused by these waves (for the wave IV a value of the order of 20 years was obtained). Such calculations were carried out for all four accumulation waves found at the stake network. Then these predicted periods were compared with the real periods observed by means of spectral analysis in the temporal series of snow accumulation at the stake network. For the waves I and II the predicted and observed periods were found to be the same (see column 1 in Table 6 where the observed periods are shown), whereas for waves IV and VI the expected periods (~20 and ~10² years) cannot be revealed by spectral analysis as the studied temporal stake series are too short (only about 30 years).

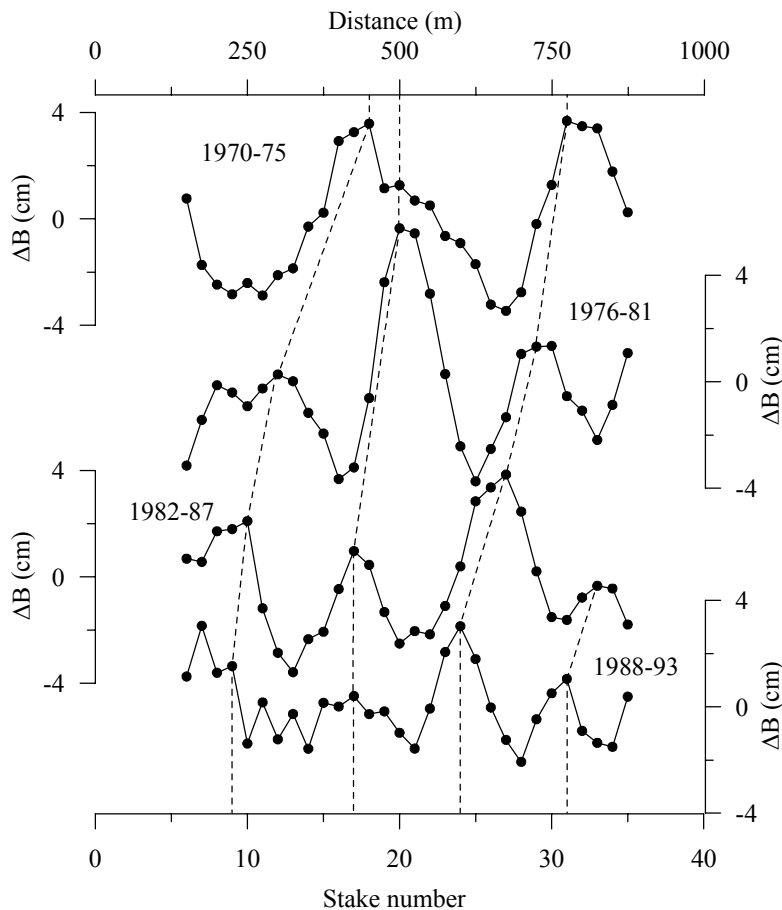


Fig. 19. Six-year snow build-up series along the NS profile of the Vostok accumulation-stake network (from [Ekaykin and others, 2002]). The series were processed with a band pass filter to extract one of the spatial accumulation waves (wave IV in Table 2). ΔB denotes the difference between the snow build-up at a given stake and the mean build-up value for the whole profile.

The periods of accumulation rate oscillations found in the time-series obtained from the pits are shown in column 3 of Table 6. It can be seen that the periods of the oscillations are the same as those at the stakes (for waves I and II) and the same as the predicted relief-related periods (for waves I, II, IV and VI). Thus, we can conclude that the periods of about 2.5, 5, 20 and of the order of 10^2 years that may be found in accumulation time-series obtained from a single point (stake, pit or ice core) at Vostok Station are likely linked to the drift of accumulation waves at the snow surface (caused by corresponding "meso-dunes" forms), rather than with change in climatic conditions.

In the accumulation time series derived from stakes and pits two oscillations were also found that do not have their counterparts in the spatial accumulation series. The periods of them are about 10 and 50–60 years (waves III and V in Table 6). The first oscillation is also revealed in the stacked (from 8 pits) time-series of snow accumulation and isotope composition, as well as

in the Vostok air temperature record (discussed in the next chapter), which confirms its climatic nature. This oscillation might be associated with the well-known solar activity cycle [Petrov, 1975]. The second cycle (50–60 years) is observed in the stacked series of snow accumulation and isotope composition over the last 200 years (see Chapter VI) and also likely has climatic origin. The variance of the climatic oscillations accounts for less than 10 % of the total variance of an accumulation time series obtained in a single point, as obtained by dividing the variance of temporal series of the mean build-up values from the whole stake-area by the mean variance of temporal build-up series from individual stakes.

In December 1999 a leveling of snow surface near each stake of the stake network was performed, which enabled to compare directly the spatial distribution of accumulation with the snow surface height. Figure 20*a* presents the smoothed and detrended snow surface profile measured along the NS section of the stake network. One of the main features of the profile is the general slope of about 6×10^{-4} , which implies that the mean surface height rises on about 60 cm from the southern to the northern end of the profile. Superimposed on this general slope are several types of the surface undulations (meso-dunes) whose horizontal lengths are listed in column 8 of Table 6. One can see that these undulations have similar lengths as the above spatial accumulation waves, which suggests the link between the latter and the meso-dunes.

In the Fig. 20 *a* and *b* we compared the snow surface height profile with the 2-year (1998–99) snow build-up profile, they both are smoothed with the period 125 m (note that the build-up scale is inversed). The oscillations corresponding to wave IV in Table 6 are clearly seen in the both profiles, and it is also seen that the accumulation waves are shifted comparing to the surface undulations, which is a necessary condition for the latter not to be filled up with time. However, the clarification of the mechanism of the meso-dunes existence requires special investigations, which is behind the scope of the present work. We just note here the link between snow relief forms on the one hand, and the spatial and temporal oscillations of snow accumulation, on the other, which is important for our study.

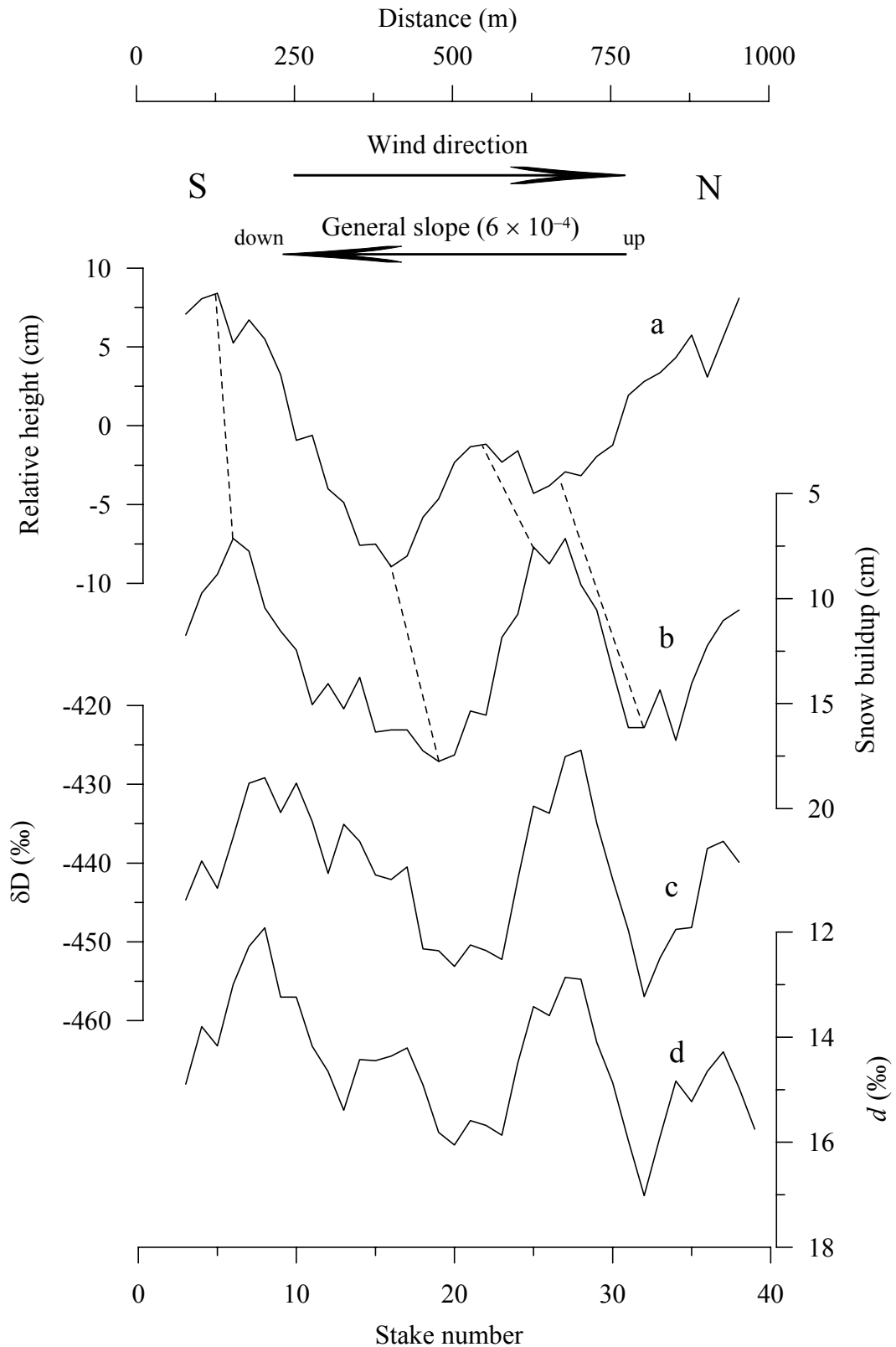


Fig. 20. The spatial profiles of snow surface height in December 1999 (a), two-year (1998–99) snow build-up (b), δD and d profiles (c and d) measured in 10-cm samples (from [Ekaykin et al., 2002]). All the profiles represent 125-m running means. The surface height profile is detrended. Note that the snow build-up and d scales are inverted.

IV.3. Relief-related oscillations in temporal isotope series

Fig. 20c presents the profile of isotope composition of the upper 10 cm of snow thickness smoothed with the period of 125 m. One can see a remarkable correlation ($r = -0.68$) between this profile and that of accumulation (Fig. 20b), the δD values being more negative when the accumulation is higher. This implies that not only the snow accumulation rate, but also snow isotope composition (and possibly other physical and chemical properties of snow) may be subjected to the spatial redistribution due to wind activity. The slope between spatial changes of surface snow isotope composition and spatial changes of annual snow build-up is -3.45 ‰/cm .

Since the annual snow accumulation at Vostok is small (about 7 cm) with respect to the magnitude of snow relief (15–20 cm), the pattern of the spatial distribution of snow precipitated during a given year is spotty, discrete, needless to say about snow precipitated during a given season. It means that at any moment of time we can find at the surface around Vostok Station snow layers formed in different recent seasons and even years. Since the properties of snow precipitation, as well as weather conditions, are different in different seasons, one can expect non-equal intensity in spatial re-distribution of summer and winter precipitation. In particular, we assume that winter snow is more easily redistributed (because of smaller snow particles and higher wind speed during winter) than summer snow. If it is true, winter snow would preferentially deposit in less windy locations (for example, behind a snow barrier, or where the surface slope is less steep). Since higher accumulation is associated with less windy locations, too, the above mechanism would produce a negative correlation between spatial distribution of snow accumulation and δD . A similar process (which is commonly called "wind scouring") has been described by Fisher and others (1983) to explain a non-climatic trend in the $\delta^{18}\text{O}$ ice-core record from Ellesmere Island, Canada.

The above-described mechanisms of wind scouring observed at Vostok (provided that they take place in reality) must also be important for all the characteristics of snow (such as chemical composition, dust concentration, etc.) that are subjected to large seasonal changes.

The additional indirect confirmation of the proposed mechanism for the inhomogeneous redistribution of precipitation formed in different seasons comes from the observed strong negative correlation between the profiles of δD and deuterium excess d (see Figure 20c and 20d, note that the d scale is inversed), if we take into account that the d seasonal cycle of in central Antarctica is inverse of that of δD (see next chapter).

Another possible explication of the observed spatial relationship between δ and accumulation is that in a location with comparatively small rate of snow build-up an annual snow layer spends comparatively more time near the snow surface where post-depositional processes are the most intense. As a result, isotope composition of this layer will be comparatively increased [Waddington et al., 2002]. Negative correlation between δD and d (Figure 20) is not in conflict with this assumption, taking into account that the condensation of the water vapor on the snow surface during the isotope exchange at the snow-air interface takes place in non-equilibrium conditions [Waddington et al., 2002].

As for the accumulation rate, we can expect that the spatial variations of δD of snow will be reflected in the time-series of isotope composition. This is simply because spatial anomalies of δD are attached to the spatial waves of accumulation (as shown at Figure 20) and their drift with time (as shown for accumulation at Figure 19) produces temporal oscillations of δD in a single point. This supposition is preliminary confirmed by the fact that the absolute magnitude (69 ‰) and the variance (278 ‰²) of spatial δD variations are almost the same as those of vertical δD profiles. It is interesting to note that all the three types of meso-dunes (waves I, II and IV in Table 6) have corresponding spatial waves of snow accumulation and corresponding anomalies in spatial δD distribution.

The periods of temporal oscillations of δD obtained from spectral analysis of δD time-series from Vostok pits are shown in Table 6, column 6 (the corresponding periods in cm of snow for the vertical δD profiles are shown in column 5). The shortest oscillation (2–3 years, wave I in Table 6) is essentially weakened in δD series in comparison to corresponding oscillations in accumulation series, which may be attributed to the diffusive isotope homogenization [Johnsen et al., 2000]. The main oscillation in the vertical δD profiles with the period of about 20–30 cm (see Fig. 11 and 12) is associated with meso-dunes with the length of the order of 100 m (wave II in Table 6). The oscillations with the period of 70 cm (about 10–12 years) likely have climatic origin. Longer variations, with period of about 120 cm (~20 years) correspond to wave IV shown at Figures 19 and 20. Wave V with period of about 50 years can be easily seen by naked eye in stacked series of isotope composition and accumulation rate of snow and represents one of the most significant modes of Vostok climatic variability (see Chapters V and VI). Finally, wave VI can be clearly seen in spatial profiles of snow accumulation (Fig. 18, see also [Ekaykin et al., 1998]). The periodicity of temporal oscillations related to the latter wave varies from several hundred to first thousands years and thus cannot be studied using available time-series. However, there is an evidence of influence of this wave on secular snow accumulation rate changes at Vostok (see Chapter VI), while its influence on snow

isotope composition still has to be confirmed. We associate wave VI with well-known megadunes observed in the most of Antarctic ice sheet.

In the end of this section we should note that existence of non-climatic relief-related temporal oscillations of snow accumulation (or isotope composition) with period of N years does not imply erasing a climatic signal with shorter periods, because the mechanism of formation of this oscillations does not involve a mixing of upper snow cover down to the depth equivalent to N -year snow build-up, as it is sometimes thought. Thus, it is possible to separate climatic and relief-related variations by applying band-pass filtering techniques. Also, non-climatic oscillations can be suppressed by constructing stacked series of accumulation (isotope composition etc.) using data from several pits (stakes, cores) due to the fact that relief-related temporal variations are not generally correlated in adjacent sites located in the same area. Distinguishing between climatic and relief-related temporal oscillations is straightforward and is based on the following principles: 1) relief-related temporal variations (unlike climatic ones) always have their counterparts in the surface (spatial) profiles of studied parameter; 2) climatic variations are synchronous in all the sites located in a given area and thus are presented in stacked series of studied parameter; 3) additional (but indirect) confirmation of climatic origin of given temporal variations is their correlation with variations of an unambiguously climatic parameter (like air temperature).

Post-depositional changes of snow δD content in the past. If we assume that the observed spatial variations of the surface snow isotope composition (Fig. 20c) is entirely explained by the post-depositional effect, then we can try to estimate possible changes of the intensity of this effect in the past. Let us suppose that the present-day spatial slope of the post-depositional δD change *versus* annual snow build-up change (-3.45 ‰ cm^{-1} , as pointed out above) is the same as the corresponding temporal δD /snow build-up slope for the Vostok area. Then, past variations of the post-depositional δD changes, $\Delta\delta D_{post-depo}$ (relative to its modern value), equal to:

$$\Delta\delta D_{post-depo} = -3.45 (b - 7) = 3.45 (7 - b),$$

where 7 is the present-day mean annual snow build-up (in cm), and b is the mean snow build-up in the past.

During the LGM epoch, the annual snow build-up could be twice (or even more) as less as at the present [Petit et al., 1999]. If we take b to be 3.5 cm, then $\Delta\delta D_{post-depo}$ is equal to about 12 ‰. Thus, to account for the post-depositional effect, we should correct the LGM isotope values, measured in the Vostok ice core, by -12 ‰ , which is roughly 20–25 % of the total δD shift during the LGM–Holocene transition. In this case, the post-depositional effect alone could explain the observed discrepancy between the "classical" paleo-temperature reconstructions

(based on the interpretation of the deep ice core isotope profile) and those based on the borehole thermometry (see the review in Section I.2.2).

In reality, one should not expect that the post-depositional δD changes would increase linearly as long as accumulation rate diminishes. As soon as the latter reaches a minimum critical value, the post-depositional processes succeed to equilibrate the isotope composition of the whole annual snow layer with that of the atmospheric water vapor, so further changes of δD of snow are not possible. It should be noted, however, that the data presented at Figure 20 *b* and *c* do not suggest that this critical value is reached even for the 2-year snow build-up of 7 cm.

Probably more important factor is the past firn temperature changes. During the LGM period, the firn temperature in the summer must have been at least by several degrees colder than now, which means that the post-depositional processes must have been less intensive [Waddington et al., 2002].

Thus, the obtained value of 12 ‰ seems to be close to the upper limit for the LGM–Holocene difference of the post-depositional changes of the upper snow δD values.

IV.5. Conclusion of Chapter IV

As a result of study of spatial variability of snow isotope composition and accumulation rate we conclude that the temporal oscillations of these two parameters with periodicities of 2.5, 5 and about 20 years observed in a single point at Vostok area are mainly linked to the drift of spatial snow accumulation waves of different scales (meso-dunes). The climatic variations of δD temporal series account for about 17 % of the total variance (and only 10 % for the accumulation rate), the most of power being concentrated in 10-year and 40–50-year periodicities. A methodology of separating relief-related and climatic temporal variations is proposed. An attempt is made to estimate possible influence of the post-depositional effects on the snow deposits δD values in the past. It was shown that due to these effects, the amplitude of the LGM–Holocene shift of the ice core isotope composition may be reduced by up to 25–30 %.

V. MODERN TEMPORAL VARIABILITY OF SNOW ISOTOPE COMPOSITION

This chapter is devoted to the following topics:

- 1) Seasonal variations of isotope composition of precipitation and its relationship to seasonal air temperature changes;
- 2) Relationship between mean annual snow isotope composition values and air temperature at different atmospheric levels.

V.1. Seasonal variability of isotope composition of precipitation

At Figure 21 the mean monthly values of precipitation isotope composition are shown as measured in the samples collected during the period from December 1999 to December 2000. The δD values change from minimum level in August (-493 ‰) to maximum level in January (-405 ‰; the value of December 2000 (-401 ‰) is based on the measurement of only one sample and cannot be considered as representative) with an annual average of -453 ‰. Thus, the annual amplitude of monthly δD values is nearly 90 ‰, while the absolute magnitude of δD values measured on individual samples is as large as 165 ‰. The sample collected during the snowfall on the 1st of August 2000 has an isotope composition of -521 ‰ being probably the lightest natural water ever measured on Earth. Thus, the seasonal cycle of isotope composition of precipitation is evident (even despite the comparatively large variability of individual δ values within a given month, see Figure 21) and is consistent with the annual cycle of local air temperature (also shown at Figure 21).

The corresponding annual amplitude of mean monthly values of $\delta^{18}O$ is 13.8 ‰. This is nearly twice as large as the value reported for 1970 [Gordienko et al., 1976] equal to 7.5 ‰. Such a big difference of the amplitudes is difficult to explain by different meteorological conditions during these years. Probably, the reduced amplitude of δ in the 1970 precipitation could be due to contamination of the samples by blowing snow though there is no direct evidence for this assumption. Indeed, the amplitude of the seasonal cycle of isotope composition of blowing snow in 2000 (9.4 ‰ for $\delta^{18}O$) was significantly lower than the $\delta^{18}O$ amplitude for precipitation (Figure 21), which suggests that blowing snow is a mixture of precipitation fallen during different seasons. At the same time, the mean annual isotope composition of blowing snow (-450 ‰ for δD) is very close to that of annual mean snowfall.

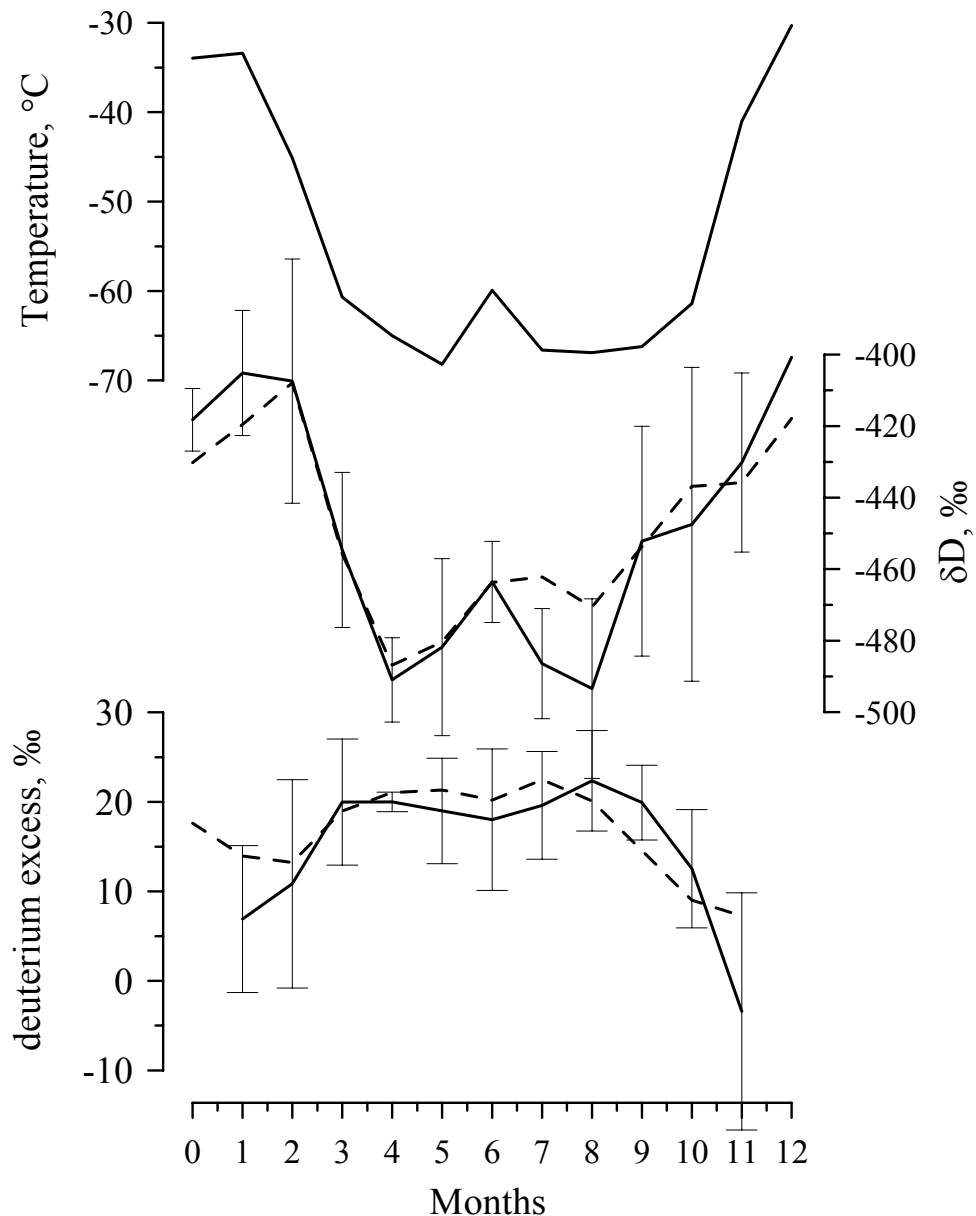


Fig. 21. Comparison of seasonal variations of isotope composition (δD and deuterium excess) of precipitation and surface air temperature (solid lines). Dashed lines denote seasonal δD and deuterium excess changes of blowing snow. Error bars represent standard deviation (1σ) of isotope composition in individual precipitation samples during given month.

Also shown at Figure 21 are seasonal changes of deuterium excess in precipitation (solid line) and blowing snow (dashed line). This parameter reveals minimum levels in summer and maximum in winter, being in anti-phase with δD and air temperature. This behavior is likely related to seasonal changes of the moisture source conditions. It is assumed that summer minimum of deuterium excess in precipitation is due to southward shift of the major moisture source [Ciais et al., 1995; Delmotte, 1997; Delmotte et al., 2000; Masson-Delmotte et al., in press; Vimeux, 1999].

An interesting feature of the annual changes of deuterium excess is the negative d value in November 2000 (-3 ‰). In individual samples of precipitation, negative d values are observed three times: 22nd of February (-3 ‰), 7th of November (-13 ‰) and 5th of December (-23 ‰). It is not usual to meet negative excess values when dealing with samples of Antarctic snow, so this issue requires more detailed discussion. One can think about several possible explanations of this phenomenon:

1) Errors of measurement. As mentioned in Section II.3.1, the accuracy of d in the discussed samples is rather low (± 1.6 ‰), which makes contribution to the observed variability of the deuterium excess values. But, obviously, this cannot explain the negative d values of 07/11/2000 and 05/12/2000 samples.

2) Alteration of the isotope composition of a sample during its evaporation (due to its small volume). Indeed, many of the samples are very small in volume (it mainly concerns precipitation samples and to less extent blowing snow samples), so they are more subjected to changes in their isotope values during storing, preparing sample for measurement, etc. In case of non-equilibrium evaporation of a sample, the deuterium excess of remaining water would become negative. However, there are several small samples of winter precipitation, and no one of them has negative d . Moreover, the 22/02/2000 sample is not small, but still it reveals negative value of excess. Thus, we have to exclude sample volume as possible factor responsible for d value disturbance.

3) Isotopic exchange of snow with air. Snow can remain in the collector comparatively long time (up to few days) until the sample is taken, which can lead to substantial isotope exchange of the snow with ambient air under the relatively warm summer conditions. As mentioned in Section I.1.3, during the night-time in summer months sublimation of water vapor exceeds evaporation likely resulting in decreasing the d value of the sample, provided that sublimation goes on in non-equilibrium, i.e., air is supersaturated in water vapor relatively to ice. The same process is probably takes place at the snow-surface interface, because two samples of blowing snow demonstrate negative values of deuterium excess, too.

4) Source conditions. We still do not exclude a possibility that the observed negative summer d in Vostok precipitation is a signature of the source conditions. Indeed, both simple isotope models and GCMs predict negative deuterium excess for the SST of $\leq 5-10$ °C and relative air humidity ≥ 85 % [Ciais et al., 1995; Delmotte et al., 2000; Petit et al., 1991; Vimeux, 1999], which corresponds to the conditions met at 50–60 °S of the Southern Ocean. One should also keep in mind that models are tuned to represent mean (over at least several years) conditions and are probably not able to reproduce isotope composition of single precipitation events. It is

worth mentioning that negative d values were obtained for some samples from Siple Dome ice core [T. Popp, personal communication, 2002] and also in few samples along the Mirny–Vostok (not published) and Patriot Hills – South Pole traverses [Qin Dahe et al., 1994].

The correlation coefficient between the mean monthly values of δD and surface air temperature (T_S)[§] for the period December 1999 – November 2000 is significant and equal to 0.89 ± 0.14 , corresponding coefficient of regression being $2.12 \pm 0.35 \text{ ‰ } ^\circ\text{C}^{-1}$. This value is considerably lower than $6 \text{ ‰ } ^\circ\text{C}^{-1}$ predicted by simple isotope models. The difference could be explained by significant change of moisture source between summer and winter periods suggested by strong intra-annual changes of deuterium excess values (Fig. 21). Another reason for the observed discrepancy is considerable seasonal changes of local conditions, in particular, inversion strength (due to changes in radiation and heat balance), which governs relationship between condensation temperature (nearly equal to inversion top temperature, as discussed in the previous chapter) and surface temperature. Indeed, if we take data only for the period from March to October (without September that shows too high value of δD , see Fig. 21), then $\delta D/T_S$ slope is equal to $4.60 \pm 1.35 \text{ ‰ } ^\circ\text{C}^{-1}$.

It is not possible to determine relationship between δD in precipitation and corresponding inversion temperature, because in 2000 the balloon-sounding observations have not already been carried out. Instead, we took mean monthly values of T_i for the period 1963–1991 keeping in mind comparatively low inter-annual variability of this parameter ($\pm 0.6 \text{ } ^\circ\text{C}$, see Chapter III). The $\delta D/T_i$ slope is equal to $6.2 \pm 1.1 \text{ ‰ } ^\circ\text{C}^{-1}$. Again, this value is less than $9 \text{ ‰ } ^\circ\text{C}^{-1}$ given by Rayleigh model. The corresponding slope for the March–October period is $9 \pm 4 \text{ ‰ } ^\circ\text{C}^{-1}$ (valid at the confidence level of 90 %).

V.2. Temporal variability of isotope composition and snow accumulation rate in the vicinity of Vostok Station over the last 50 years

The main feature of the vertical profiles of snow isotope composition (δ) is the regular oscillations (Fig. 11 and 12, and Table 6) with the total amplitude of about 60–80 ‰ for δD (that is, about 70–90% of the seasonal change of isotope composition of precipitation at Vostok, see Section V.1 and Figure 21). Similar oscillations have been discovered in the vicinity of old Dome C Station [Petit and others, 1982] that resembles Vostok in its geographical and meteorological conditions (mean annual accumulation is 3.2 g cm^{-2}). Note that the value 60–80

[§] See Annex A for the list of basic notations and abbreviations used in this work

‰ of the amplitude is larger than the magnitude of δD change during LGM-Holocene transition (about 50–55 ‰ [Petit and others, 1999]).

As shown in Chapter IV, these oscillations are mainly related to the drift of different types of snow surface relief forms (micro-relief and meso-dunes) rather than to climatic variations. The signal-to-noise ratio [Fisher et al., 1985], determined from the mean correlation coefficient of the δ time-series from 8 individual pits, is 0.2, which suggests that "stratigraphic noise" accounts for at least 83 % of the total variance of snow isotope composition in a single point. This noise linked with the extremely low accumulation rate was previously found when investigating snow accumulation at the Vostok stake network (see Section IV.2). To reduce the noise, we constructed the stacked δD time-series for all the eight pits. The new series, hereafter referred to as "*pit*" (1943–1998), is shown at Fig. 22 together with the δ series from individual pits.

Based on the results of Chapter IV, we can assume that the resolution of δ time-series obtained in a single point at Vostok in terms of possible climate signal is not better than 20 years. This period is needed to eliminate the relief-related variations linked with the largest meso-dunes (wave IV in Table 6). For the stacked "*pit*" series the corresponding period was estimated to be 7 years [Ekaykin et al., 2002]. Further smoothing decreases the variance of the series insignificantly and does not increase the signal-to-noise ratio.

The smoothed "*pit*" series is shown at Figure 22 by a thin line. The isotope composition of snow changed significantly during the last 60 years with the minimum in 1953–1964 and maximum in the 1980s followed by a decrease during the last 10 years. The mean isotope composition of snow as obtained from pit studies is -440 ‰ with an inter-annual variability of ± 9 ‰ (1σ). The lowest value of the "*pit*" series (-453 ‰) observed in 1954 is coincidentally equal to the mean precipitation isotope composition in 2000 (Section V.1). The isotope composition of the snow layer dated by 2000 and sampled in December 2001 in 48 mini-pits around Vostok is -451 ± 2 ‰, which is nearly equal to the isotope composition of the precipitation collected in 2000. The given values suggest that by the end of the last decade the δ of precipitation nearly reached the absolute 60-year minimum value following the temperature cooling observed during the same period.

In Table 7 the correlation coefficients between smoothed "*pit*" series and air temperature at different atmospheric levels are shown.

The correlation (r) between snow isotope composition and surface air temperature is statistically significant (with the probability of 95 %) and equal to nearly 0.6. Thus, local temperature at Vostok accounts for about 40 % of the isotope inter-annual variability. The

corresponding regression coefficient is $17 \pm 4 \text{ ‰ } ^\circ\text{C}^{-1}$. This value is almost 3 times larger than the $\delta\text{D}/T_s$ slope predicted by simple isotope model. Similar value ($20.3 \text{ ‰ } ^\circ\text{C}^{-1}$) was obtained by Jouzel and others (1983) for the South Pole precipitation. The reason for such large isotope-temperature slopes is not yet very well understood.

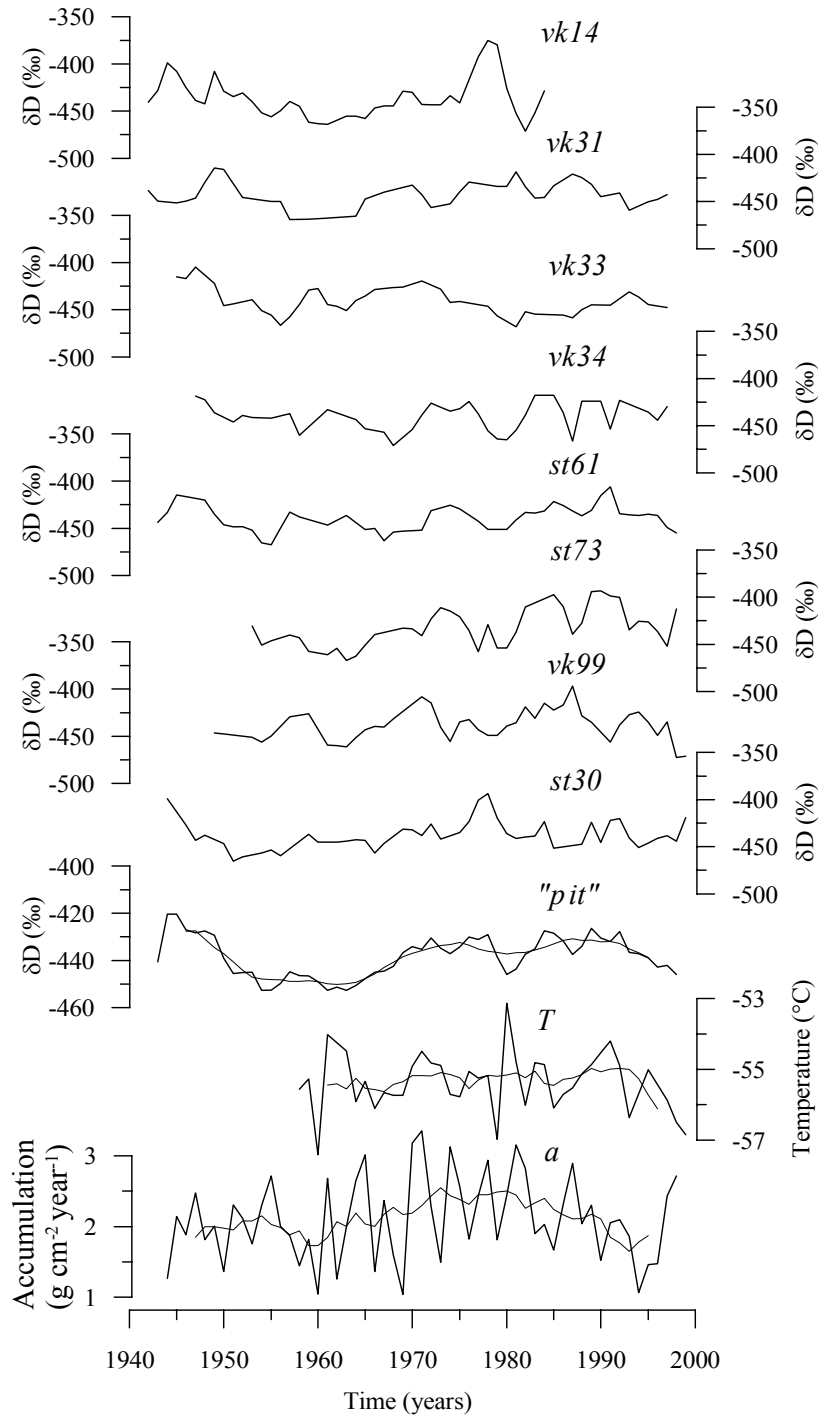


Fig. 22. The time-series obtained from shallow snow pit studies at Vostok: mean annual snow isotope composition (δD) in the individual pits *vk14*–*st30* and the stacked "pit" series for

all the eight pits; a – snow accumulation rate. T – surface air temperature as obtained from meteorological observations. The smoothed "pit", T and a series are shown by the thin lines.

Table 7. Correlation coefficients between air temperature at different atmospheric levels and stacked series of isotope composition (δD) and accumulation rate (a) of snow as obtained by studies in 8 shallow pits.

r	T_s	T_{600}	T_i	T_{500}	T_{400}
δD Vostok	0.57±0.14	0.69±0.16	0.0	0.0	-0.1
δD Vostok*	0.76±0.12	0.63±0.19	0.37±0.23	0.1	-0.36±0.23
δD South Pole	0.47	0.54	0.57	0.46	no data
a Vostok	0.15±0.17	0.16±0.22	0.21±0.22	0.44±0.20	0.39±0.21

All Vostok series are smoothed with 7-year running filter except for *, which is smoothed with 11-year filter (from [Ekaykin et al., submitted]). Statistically significant coefficients with a confidence level of 95 % are shown in bold. Values of r between δ and T at South Pole Station are given for comparison [Jouzel et al., 1983].

Surprisingly, the correlation between δD and inversion temperature is equal to zero. Assuming that the inversion layer is the one where most of precipitation forms, one could expect the highest (comparing to the other atmospheric levels) r value between δD and T_i , as it's the case for South Pole (see Table 7). At Figure 23a series of δD in Vostok snow and T_i are shown for the period 1963–1991.

One can see that both series can be decomposed into two components: a trend that is similar for both parameters (this trend is actually a part of 40–50-year oscillation seen at Figure 22; see also next Chapter) and a short-term component with the period of about 10 years that is in anti-phase for the δD and T_i . Obviously, the combination of these components produces zero correlation between the resultant series of isotope composition and temperature. The negative relationship between 10-year oscillations of δD and T_i means that another parameter rather than local temperature is responsible for the isotope variations. The fact that deuterium excess series, shown at Figure 23b, is well correlated T_i (and anticorrelated with δD) at the 10-year time scale suggests the influence of source conditions as a primary cause of the observed phenomenon.

We have made an attempt to correct the deuterium content data for the source conditions using the data on d and following the method developed by Vimeux et al. (2002), see Equation 16b. The third term of the equation was taken as zero since we assume constant isotope

composition of the water in the source region for the time period of interest. For our calculations we took the values for the coefficients as listed in Table 2.

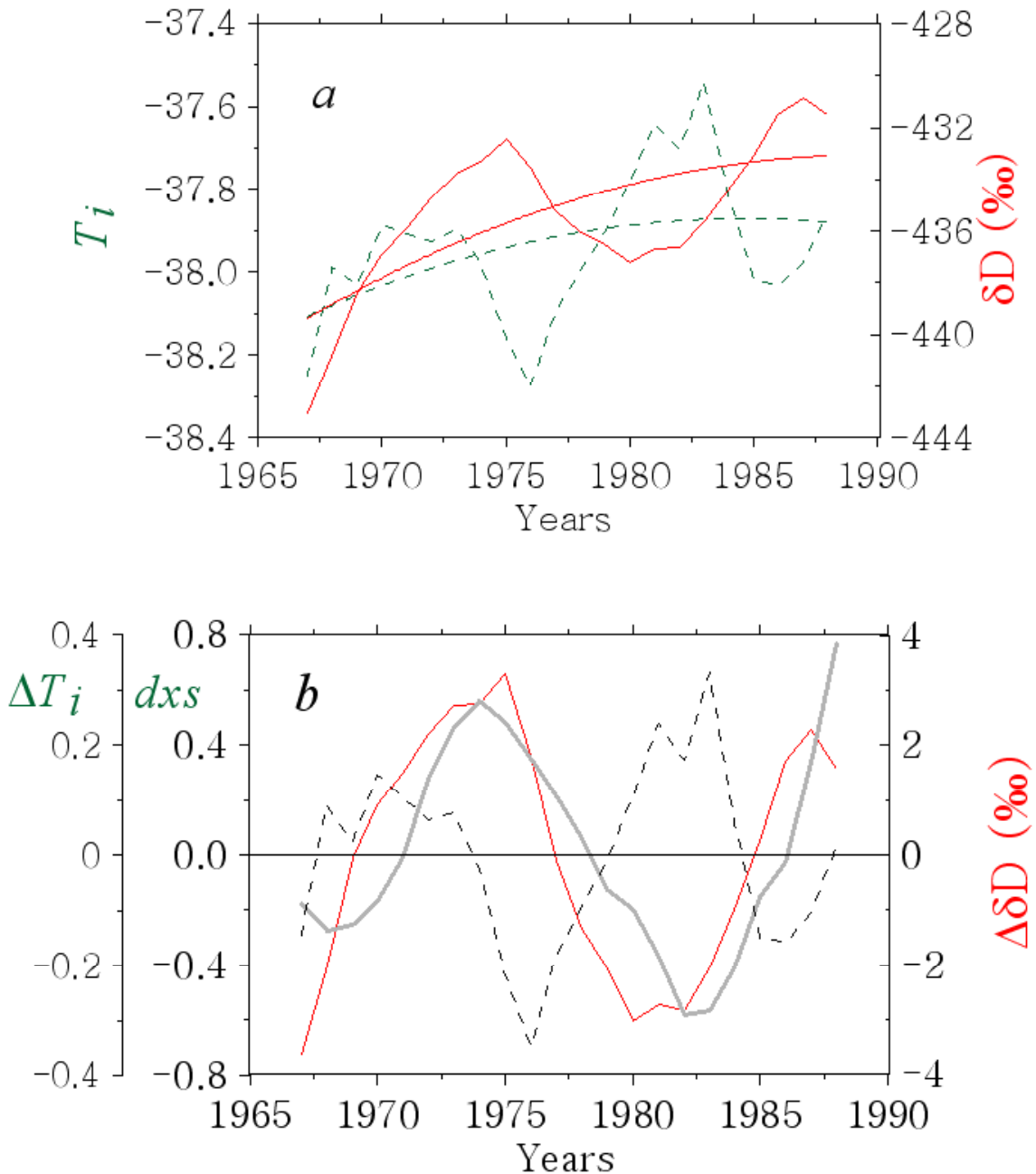


Figure 23: *a* – series of deuterium content in Vostok snow (solid line) and inversion temperature (dashed) for the period 1963–1991, smoothed with a 7-year running filter, and their trends; *b* – residuals (series value minus trend value) of the two parameters shown in panel *a* and of deuterium excess of snow (thick grey solid line).

As a result, the calculated site condensation temperature showed negative correlation (-0.34 ± 0.21) with Vostok's inversion temperature. This happened simply because by applying the correction we amplified short-term (10-year) variations of δD , since δD and d are in phase

one with another. Let us consider whether it is possible theoretically that local temperature and isotope composition of precipitation are in opposition. For this, we take equation 16b without the third term and try to figure out which coefficients may be responsible for the observed relationship between T_i (that is believed to be an approximation of T_C) and δD :

$$\Delta T_C = (\gamma_{SST}/\Omega)\Delta d + (\beta_{SST}/\Omega)\Delta\delta D \quad (18)$$

As shown at Figure 23b, Δd and $\Delta\delta D$ are in phase one with another, and both of them are in opposition with the observed variations of ΔT_C (inversion temperature). This situation is possible when either γ_{SST} or β_{SST} (or both) coefficients are negative, which would be very difficult to explain. Another (and the last) option is that Ω parameter, which is an arithmetic combination of the four coefficients ($\Omega = \gamma_C\beta_{SST} - \gamma_{SST}\beta_C$, see Table 2 for the coefficients definitions and values) is negative. Tuning these four factors (γ_C , β_{SST} , γ_{SST} and β_C), it is possible to make Ω negative keeping all of them positive. In particular, negative Ω means that $\gamma_C\beta_{SST}$ is less than $\gamma_{SST}\beta_C$. This would require comparatively large changes of some (or all) of these coefficients comparing to their values obtained from simple isotope models (see Table 2). For example, these models suggest that $\gamma_C\beta_{SST}$ equals to $9.23 (\% \text{ } ^\circ\text{C}^{-1})^2$, while $\gamma_{SST}\beta_C$ is $1.85 (\% \text{ } ^\circ\text{C}^{-1})^2$. In order to make $\gamma_C\beta_{SST}$ be less than $\gamma_{SST}\beta_C$, one should:

diminish the value of γ_C (which implies weaker influence of local temperature on the isotopic composition of precipitation. This is in agreement with the fact that the slope between the seasonal changes of δD and T_i at Vostok is about 30 % less than the corresponding slope from an isotope model);

and/or diminish the value β_{SST} (which means weaker role of the source conditions on the deuterium excess of snow);

and/or increase the value of γ_{SST} (which implies stronger role of the source conditions in the changes of the δD values at Vostok);

and/or increase the value of β_C (which means stronger influence of the local temperature on the deuterium excess of the snow precipitation).

Thus, we conclude that short-term (about 10 years) variations of isotope composition of snow at Vostok seem to be dominated by source rather than local conditions. The involved mechanisms are likely characterized by substantial deviations from mean climatic conditions. At the same time, we should note that the correction of the δD series for the source conditions does not change correlation between δD and T_S . So, in further study the uncorrected δD deuterium content series is used.

To reduce the influence of these 10-year variations we smoothed both δD and temperature series by 11-year running filter and showed the correlation coefficients between

them in Table 7. The r value of $\delta D/T_i$ relationship is now positive (0.37 ± 0.23) though statistically insignificant due to too short series of temperature. The corresponding regression coefficient is $5.4 \pm 3.4 \text{ ‰ } ^\circ\text{C}^{-1}$, which is less than $9 \text{ ‰ } ^\circ\text{C}^{-1}$ even taking into account its large uncertainty. The correlation coefficient of δD with T_S has become 0.76 ± 0.12 with the slope of $28 \pm 4.5 \text{ ‰ } ^\circ\text{C}^{-1}$.

Generally, the $\delta D/T$ correlation coefficients listed in Table 7 are explained by the relative intensity of long-term temperature changes during the period 1963–1991. The trend of temperature is stronger near the surface than in free atmosphere, which explains the highest correlation observed for T_S . The trend is the weakest for the 500 hPa level, while for the 400 hPa the trend's sign is opposite, which is reflected in negative r between δD and T_{400} .

The correlation coefficients between snow accumulation rate and air temperature is positive for any atmospheric level and vary between 0.15 and 0.44 (all of them are insignificant).

The slope between temperature changes at the top of inversion and near the surface (parameter C_i , see equation 13) is equal to 0.7–0.8 and is slightly dependent on the period of smoothing of the both series. This value is close to the corresponding present-day spatial coefficient of 0.67 reported by Jouzel and Merlivat (1984) and widely used in paleoclimatological isotope reconstructions. The C_i value for the seasonal changes of T_i and T_S at Vostok is 0.32 ± 0.02 , which is two times less than the above mentioned spatial slope. This low value is explained to large extent by the summer disintegration of the inversion layer. In this case, C_i is approximately equal to $1 - \Delta T_i / \Delta T_S$, where ΔT_i is inversion–surface temperature difference in winter (about $25 \text{ }^\circ\text{C}$) and ΔT_S is magnitude of seasonal surface temperature changes (about $35 \text{ }^\circ\text{C}$). If only winter months are taken (April–September), then the C_i is equal to 0.70 ± 0.25 , which is close to the corresponding inter-annual and spatial slopes.

V.3. The deuterium content – temperature slopes

In this section we make an overview (Table 8) of different available estimations of the δD –temperature slopes, mentioned in this work (both taken from literature and obtained during the present study).

The results of our study, in general, suggest that the present-day empirical temporal slopes between the precipitation isotope composition and the air temperature (both at the condensation level and near the surface) tend to be less than the theoretical ones (predicted by simple isotope models). The only exception is the inter-annual $d\delta D/dT_S$ slope, which is about 3 times higher than the theoretical one. The origin of this phenomenon is not very well understood.

Table 8. Regression coefficients between spatial and temporal (seasonal and inter-annual) changes of snow isotope composition and air temperature ($\text{‰ } ^\circ\text{C}^{-1}$). Also shown are the spatial and temporal slopes between inversion, condensation and surface air temperatures.

	$C = d\delta D/dT_S$	$C_T = d\delta D/dT_C$	$C_i = dT_i/dT_S$	$C_C = dT_C/dT_S$	Reference
Spatial					
DDU-DC	6.04				LM 79
Antarctica			0.67		JM 84
Mir-Kom	6.4±0.2				E 01
Mir-Vos			0.60	0.45–0.52	this work
theoretical		9			JM 84
Seasonal					
Vostok	2.12±0.35	6.2±1.1	0.32±0.02	0.32±0.02	this work
Vos, winter	4.6±1.35	9±4	0.70±0.25	0.70±0.25	this work
Inter-annual					
Vostok	17–28	5.4±3.4	0.7–0.8	0.7–0.8	this work
South Pole	20.3	17			J 83
LGM – Holocene transition					
Vostok	6	9	0.67	0.67	P 99
Vostok	4.5±0.5	7.5±1.5	0.6±0.1	0.6±0.1	S 98

T_S is surface air temperature; T_i is the temperature at the top of the surface inversion layer; T_C is the overall condensation temperature at which the formation of the precipitation takes place. For the area of Vostok, it is supposed that T_C equals to T_i within the uncertainty of their values (see Section III.2). References: LM 79 = Lorius, Merlivat, 1979; JM 84 = Jouzel, Merlivat, 1984; E 01 = Ekaykin et al., 2001; J 83 = Jouzel et al., 1983; P 99 = Petit et al., 1999; S 98 = Salamatin et al., 1998. Abbreviations: DDU-DC = Dumont d'Hurville – Dome C; Mir-Kom = Mirny – Komsomolskaya; Mir-Vos = Mirny - Vostok. Statistically significant coefficients are in bold.

From our point of view, the disagreement between the "classical" isotope calibration (based on the present-day spatial isotope-temperature relationship [Jouzel, Merlivat, 1984; Petit et al., 1999]) and the alternative one (based on the borehole thermometry [Salamatin et al., 1998]) could be overcome, if we take into account the following circumstances:

1) The C_i coefficient, used by isotope models (0.67) seems to be overestimated. It is not clear, which of the available values of this coefficient (present-day spatial, seasonal and inter-annual) is the best approximation for the inversion strength changes during the LGM–Holocene transition, but in any event the value of 0.67 lies at the upper limit of a possible range (0.32–0.70). Also, as shown in Section I.3.4, the use of the spatial C_i (inversion *vs* surface air temperature, equal to 0.67) instead of C_C (condensation temperature *vs* surface air temperature, equal to 0.45–0.52) is incorrect and may lead to a relative error of up to 30 %.

2) As shown in Section IV.3 (sub-section "Post-depositional changes of δD content of snow in the past"), the magnitude of the δD shift during the LGM–Holocene transition is probably weakened (up to 30 %) due to stronger post-depositional processes in the LGM epoch. If one would apply the corresponding correction to the isotope profile, than the amplitude of the LGM–Holocene temperature change produced by the classical approach would be stronger even without changing the isotope-temperature slopes.

3) Probably, the best way of tuning a simple isotope model, when trying to obtain the isotope-temperature slope for the glacial-interglacial transition, is to use the seasonal isotope-temperature relationship for the winter months. The use of the full seasonal cycle seems to be incorrect because of too different meteorological conditions in summer (positive radiation balance) and winter, which does not realistically represent the LGM–Holocene transition. The same concerns (even to greater extent) the use of the spatial isotope-temperature slopes. The use of the inter-annual isotope-temperature relationships is not possible, too, because of too low amplitude of the temperature changes in central Antarctica.

The refining of the isotope model is far beyond the scope of the present study. Special investigations will be carried out in the future to complete this task.

V.4. Short-term variations of isotope composition in deep ice cores from Vostok

According to Johnsen (1977) and Johnsen and others (2000), all the oscillations in vertical δD profile with the period of about 20 cm of ice equivalent are completely erased during the firnification process. For the upper 3 m of snow thickness at Vostok (average density is 0.36 g cm^{-3}) this critical period corresponds to about 50 cm of snow. Comparing this value with the data shown in Table 6, we come to the conclusion that the shortest δD oscillations to be expected in the deep ice-core isotope record is wave III with the period of about 12 years (assuming no changes in spectral characteristics of the δD series in the past).

At Fig. 24 the results of δD measurements in the deep Vostok ice core samples NN 123, 183, 1819 and 2136 are shown. The ages of ice are 5, 8, 127 and 150 kyrs, correspondingly. The deep profiles are compared to the δD profile from the *st61* pit smoothed with the period 25 cm. Note that the *st61*, N 123 and N 183 profiles (Holocene) has the same mean δD value, while N 1819 (last interglacial) is isotopically heavier and N 2136 (close to the penultimate glacial maximum) is isotopically lighter. The profiles are reduced to a common length scale, so that an average annual layer in each profile has the same thickness.

The profiles shown at Fig. 24 have obvious common features. In particular, the profiles *st61*, N 123 and N 1819 have similar period of oscillations (about 10–12 years), though the amplitude of δD changes in N 1819 is much weaker than in *st61*, probably due to the diffusive smoothing. For profile N 183 the wavelength (≥ 57 cm of ice, which corresponds to about 25 years) is close to those of wave IV (see Table 6), though the length of the profile does not allow to be sure in this result. In profile N 2136 the period of δD variations is about 44 years, which is similar to the wave V (Table 6).

In general, detailed isotope profiles from deep ice cores reveal the same oscillations as the present-day vertical isotope profiles, which suggest that the origin of these oscillations may be similar. Thus, the results obtained by the study of snow accumulated at Vostok Station during the last 50 years give a clue to understanding of the δD variations in ice deposited in this area in the remote past.

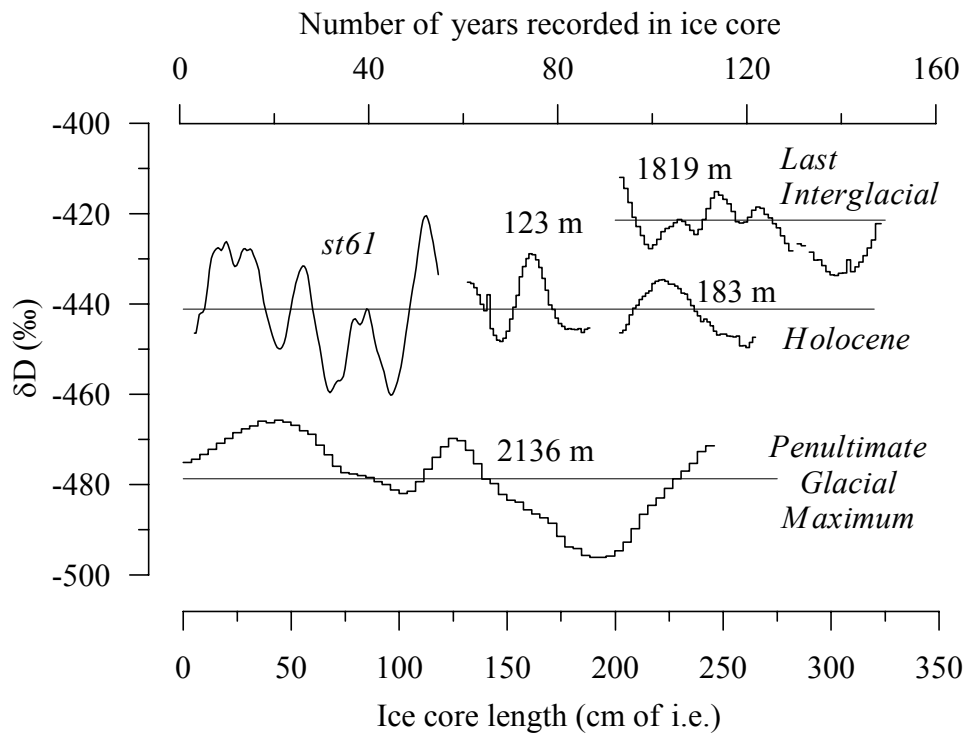


Fig. 24. Detailed isotope (δD) profiles from the deep Vostok ice core increments N 123 (5 kyrs BP), N 183 (8 kyrs BP), N 1819 (127 kyrs BP) and N 2136 (150 kyrs BP) compared with the δD profile from the *st61* pit (from [Ekaykin et al., 2002]). The *st61* pit profile is smoothed with the period 25 cm. All the profiles were reduced to a common time scale by correcting for annual layer thinning, density and accumulation rate differences. The resulting uniform time scale is given in the upper part of the figure.

V.5. Conclusion of Chapter V

The magnitude of seasonal changes of δD in precipitation is about 90 %. Monthly values of δD are correlated with surface and inversion air temperature with coefficient of regression, correspondingly, 2.12 ± 0.35 and 6.2 ± 1.1 ‰ °C⁻¹. Seasonal variations of deuterium excess are opposite to those of temperature and δD and are likely related to changes of Vostok moisture source conditions. Summer minimum of d is also probably deepened by post-depositional processes in the upper snow thickness.

The inter-annual changes of snow isotope composition are well correlated to the surface air temperature, but the coefficient of regression is very sensitive to the period of smoothing of the studied series (for periods of 7 and 11 years it varies from 17 to 28 ‰ °C⁻¹). This correlation is mainly determined by slow (50-year) variations, while at short-term time-scale (about 10 years) the isotope composition of snow at Vostok seems to be primarily governed by source rather than local conditions.

The slope between inter-annual changes of inversion and surface temperature is equal to 0.7–0.8, which is close to the value of 0.67 currently used in the paleo-climatic studies.

Detailed isotope profiles from deep ice cores reveal the same oscillations as the present-day vertical isotope profiles, which suggest their similar origin.

VI. CHANGES IN TEMPERATURE AND SNOW ACCUMULATION RATE AT VOSTOK STATION OVER THE PAST 200 YEARS

This chapter is addressed to the following questions:

- 1) What are the main features of the air temperature (T) and snow accumulation rate (a) changes at Vostok over the last 200 years?
- 2) Are there a and T trends related to the global warming?
- 3) Do the mega-dunes have any influence on the secular trends of these two parameters at Vostok?
- 4) What are the 200-year trends of accumulation and temperature at other East Antarctic sites?

VI.1. Series of isotope composition and snow accumulation rate from deep pits

Based on the results of the stratigraphic and geochemical observations in three deep pits (see Section II.2 and Figure 12) temporal series of snow isotope composition and accumulation rate have been reconstructed for the last 225 years (1774–1999). Stacked δD and a series are presented at Figure 25*b*. One should note a high level of stratigraphic noise in the variability of both parameters. The required smoothing to suppress sufficiently this noise was estimated in Section V.2 to be 7 years for both δD and a . Note that for δD series the contribution of high-frequency noise is less than for accumulation due to diffusive isotope smoothing [Fisher et al., 1985; Johnsen et al., 2000].

Results for the period 1943–1999 are shown at Figure 25*a* together with the following data (all the series are smoothed by 7-year running filter):

- 1) mean annual surface air temperature (T_S) as obtained from instrumental observations at Vostok meteorological station in 1958–1999;
- 2) stacked δD and a series from 8 shallow pits for the period 1943–1998;
- 3) snow accumulation rate series as obtained from snow build-up measurements at the Vostok stake network in 1970–1999.

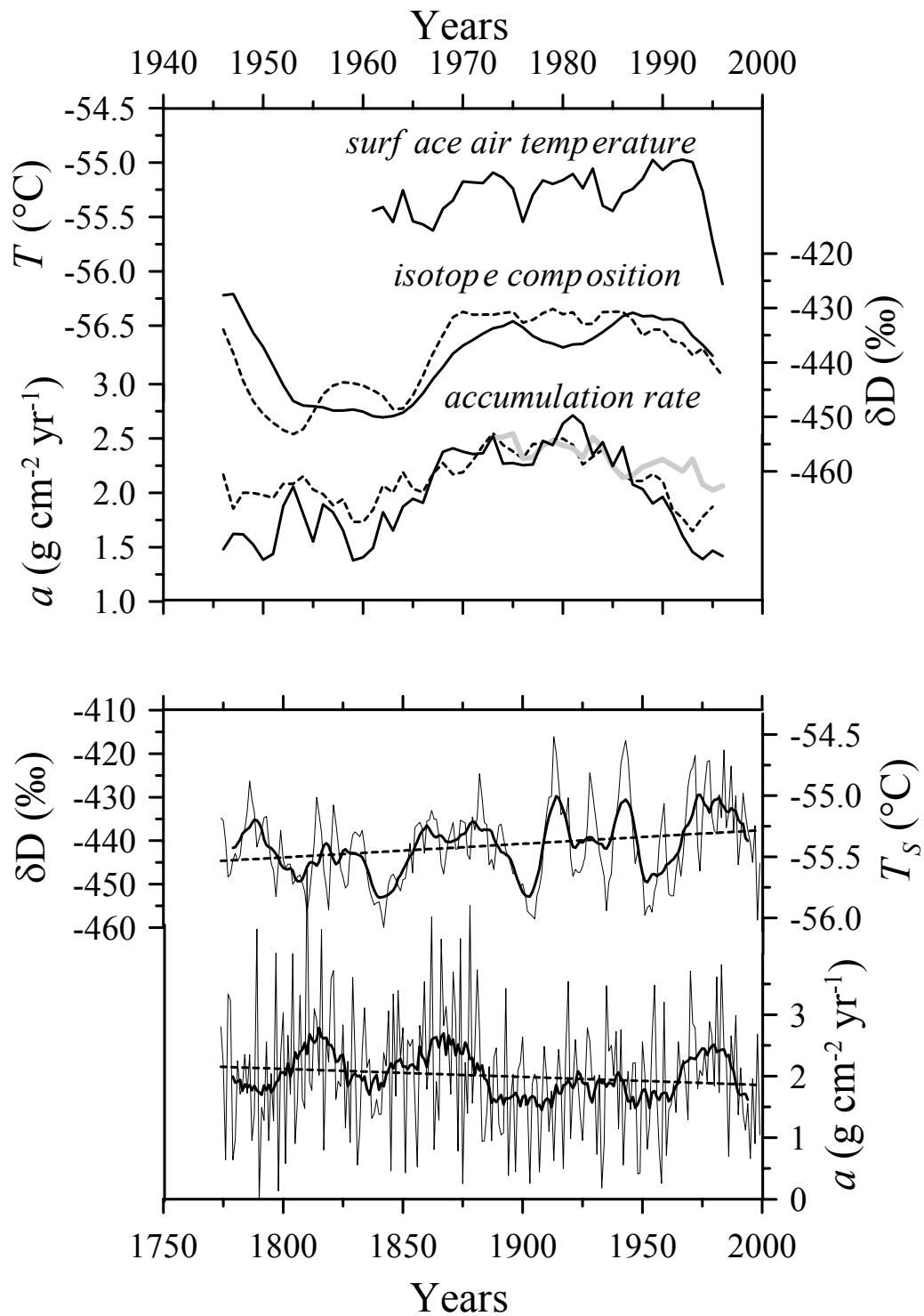


Fig. 25. a) Comparison of deep pit series to the data on δD , a and T available for Vostok: surface air temperature, a and δD from 8 shallow pits (dashed line) and a from the stake network (thick grey line). All the series are smoothed with a 7-year running filter.

b) Stacked series of δD and a from deep pits. Thin lines are original series, while the thick ones represent 11-year running means. Surface air temperature scale (T_s) was calculated according to the $\delta\text{D}/T_s$ slope of $28 \text{ ‰ } ^{\circ}\text{C}^{-1}$ (see Section V.2). From [Ekaykin et al., submitted].

All the six series reveal similar features, which result in the fact that the correlation coefficients between all the parameters are positive and almost all are significant. Values of δD , T and a are decreasing in 1950s with the minimum about 1960 and then increasing until a maximum in the 1970s. The only exception is the series of temperature with the maximum around 1990. These features are likely common for the whole East Antarctica (see, for example, [Morgan et al., 1991; Pourchet et al., 1983]). According to M. Pourchet with co-authors (1983), who obtained snow accumulation values in 14 sites by snow β -radioactivity measurements, the mean snow accumulation rate in East Antarctica in 1965–1977 was 30 % higher than during the previous decade.

During the last 10–20 years all the series reveal a clear decrease of their values, well marked for air temperature. This agrees with cooling observed at the most Antarctic stations (except for Antarctic Peninsula) during the 1990s [Doran et al., 2002].

The observed changes are likely related with the variations of cyclonic activity in Antarctic [Enomoto, 1991; Morgan et al., 1991; Simmonds and Keay, 2000], since cyclones bring both moisture and heat to the interior of the continent.

Bold lines at Figure 25*b* represent δD and a series smoothed with a period of 11 years. This smoothing window was chosen to eliminate 10-year variations of deuterium content that are supposed to be dominated by source conditions (Section V.2). The temperature scale for the isotope series was constructed using the regression coefficient of $28 \text{ ‰ } ^\circ\text{C}^{-1}$ established for the $\delta D/T_S$ relationship for the last 40 years. According to this reconstruction, 11-year means of surface annual air temperature has changed over the past two centuries between -55 and $-56 \text{ }^\circ\text{C}$. Thus, mean present-day (1958–1999) surface air temperature value (-55.4°C) obtained as a result of meteorological observations is close to its average over the last 200 years. During the past two centuries snow accumulation rate changed between 1.5 and $3 \text{ g cm}^{-2} \text{ year}^{-1}$.

In general, the comparison of the series of isotope composition and snow accumulation rate from deep pits with other Vostok data on δD , a and T allows to conclude that the 200-year series represent climatic variability in the studied area of Antarctica over the last 40–60 years and thus are likely representative over the whole period of interest.

One can also note the opposite linear trends of the δD and a series over the last two hundred years, clearly seen at Figure 25*b*. This feature is discussed below in Section VI.3.

VI.2. 50-year cycle in changes of accumulation and isotope composition: A teleconnection between central Antarctica and tropical Pacific?

Spectral analysis of 200-year series of a and δD (not shown) revealed cycles with period of about 50–60 years, more pronounced for accumulation than for isotopes. Based on the data shown in the previous section it can be assumed that the observed variations are linked to the changes of cyclonic activity around Antarctica [Enomoto, 1991; Morgan et al., 1991]. In particular, Enomoto (1991) analyzed data on mean July sea surface pressure in 40–50 °S band over the period 1873–1981 and found a 40–60 year cycle of meridional circulation. Interestingly, accumulation changes at Vostok are in anti-phase with air pressure in the Pacific and in phase with those in the Indian ocean, which suggests that the former is likely the dominant source of moisture for the Vostok area [Averianov, 1969a; Ekaykin et al., 2001] rather than the latter [Delaygue et al., 2000].

Recently, semi-centennial variations were found in changes of sea surface temperature, intensity of zonal and meridional circulation, CO₂ concentration and biological production in the Pacific Ocean [Chavez et al., 2003]. These variations are related to the so-called Pacific Decadal Oscillation which mechanism is similar to that of El-Nino (e.g., [Trenberth, Hurrell, 1994]). The most prominently these oscillations can be seen in see-saw changes of anchovy and sardine fish abundance. The "sardine" phase of the cycle (observed from mid-seventies to the beginning of nineties of the last century) is characterized by higher SSTs in central and eastern Pacific and lower SSTs in the north-west and southern parts of the ocean. For the "anchovy" phase (observed from early 50s to middle 70s and started again in the 90s) the opposite SST distribution is typical. Captivatingly, both a and δD at Vostok are practically in phase over the last century with the PDO index (Figure 26), which suggests a relationship between these parameters. The correlation between climate series in central Antarctic and tropical Pacific implies a teleconnection between these two regions on decadal scale and indirectly supports that Pacific Ocean is the dominant moisture source for the Vostok area.

The correlation coefficient between the δD series and the PDO index for the last 100 years is 0.44 ± 0.10 (for the 11-year running mean values). Surprisingly, correlation of PDO with deuterium excess series is only -0.1. One could expect more close relationship between these two parameters since d is thought to be a proxy of source conditions. To investigate this issue, we calculated the source temperature (T_{source}) for Vostok for the last 200 years using δD and d data from deep pits and employing the method of Vimeux et al. (2002) (see equation 16a) and plotted T_{source} together with the PDO index (Figure 27).

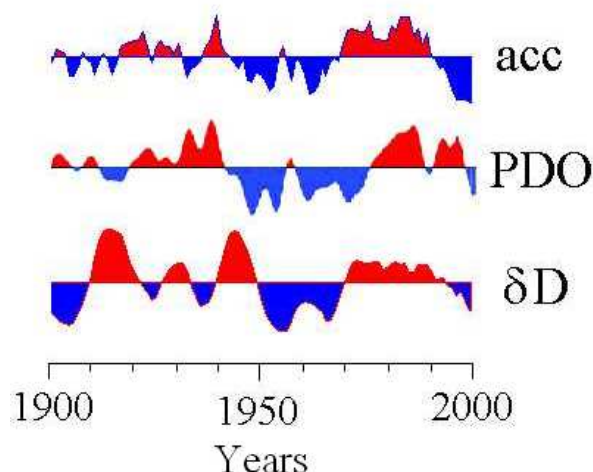


Fig. 26. Comparison of Vostok snow accumulation and isotope composition series with variations of PDO index over the period 1900–2000. All the series are de-trended and smoothed with the period of 7 years.

One can see that during the first half of XX century both parameters reveal remarkable positive correlation ($r = 0.90 \pm 0.06$). Then, in the middle 60s the sign of correlation changed and during the period 1963–1999 relation between T_{source} and PDO was negative ($r = -0.77 \pm 0.12$). The r coefficient for the whole series is 0.23 ± 0.10 . Such a behavior of correlation between these two parameters suggests a sharp change of atmospheric circulation in the 1960s which probably led to alteration of major Vostok moisture source area or pathways. This assumption is confirmed by the observed large reorganization of the tropical climate variability in 1970s, possibly linked with changes in the southern Pacific Ocean [Giese et al., 2002]. During the same period, anomalous peak of deuterium excess is observed in the Law Dome ice core record [Masson-Delmotte et al., in press] explained by intensified warm transport from low latitudes of Indian Ocean to Antarctic coast. This corresponds to the positive anomaly of Vostok moisture source temperature (Figure 27) centered around 1975. The similarity between Vostok ΔT_{source} series and Law Dome deuterium excess record (see Figure 8 from [Masson-Delmotte et al., in press]) for the period 1965–1990 suggests that since middle 1960s the contribution of Indian Ocean as Vostok moisture provider has increased relatively to the Pacific. To verify this tentative assumption thorough investigation of the involved processes are needed.

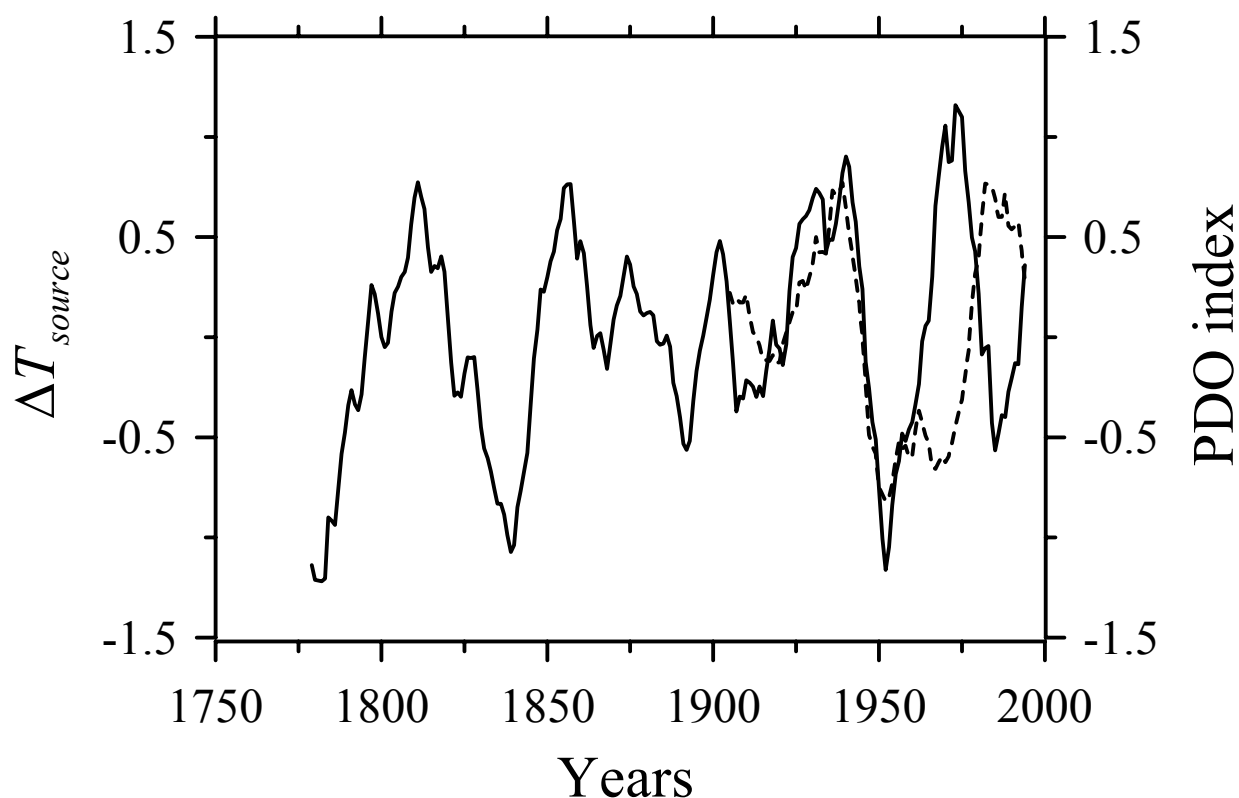


Fig. 27. Comparison of reconstructed Vostok moisture source temperature (ΔT_{source}) with the PDO index. Both series are smoothed by 11-year running filter.

Another feature of the studied series that attracts attention is the opposite sign of the linear trends of a and δD over the last 226 years (Fig. 25b). During the period of interest snow accumulation rate decreased by $-0.3 \text{ g cm}^{-2} \text{ year}^{-1}$ (20 % of the total magnitude of 11-year means) and δD increased by 7 ‰ (nearly 30 % of the total magnitude). The trend of deuterium content is statistically significant with the confidence level of 95 %. The fact that the sign of a trends in individual series is opposite for pits located only 2 km apart (see *vk10* and *st30* in Fig. 28 where accumulation and isotope series from individual pits are shown) casts doubt on climatic origin of the trends of a and δD in the stacked series. The most likely reason of the observed phenomenon should be the drift of large snow relief forms (so-called "mega-dunes") through the pit sites.

VI.3. Secular trends of accumulation and isotopes at Vostok: Climate or mega-dunes?

A number of studies (e.g., [Black and Budd, 1964; Frezzotti et al., 2002; Gow and Rowland, 1965; Whillans, 1975; Van der Veen et al., 1999]) have shown that large snow dunes observed everywhere on the surface of Antarctic ice sheet are responsible for the formation of quasi-periodical spatial snow accumulation waves with the wavelengths from 2 to 40 km. Drift

of these dunes due to wind activity causes temporal oscillations of a in a given point, the period of those oscillations being related to the dunes horizontal scales through their drift velocity. Since the above velocity is estimated to be of the order of 20–25 m yr⁻¹ [Black and Budd, 1964; Whillans, 1975], the expected period of the temporal oscillations is from several hundred to first thousands of years.

Because of these relief-related oscillations, time series of snow accumulation obtained in adjacent ice cores are poorly correlated even in centennial time scale [Van der Veen et al., 1999]. The period of smoothing required for eliminating this long-term relief-related noise in a series obtained at South Pole was estimated by Van der Veen and others (1999) to be about 3600 years.

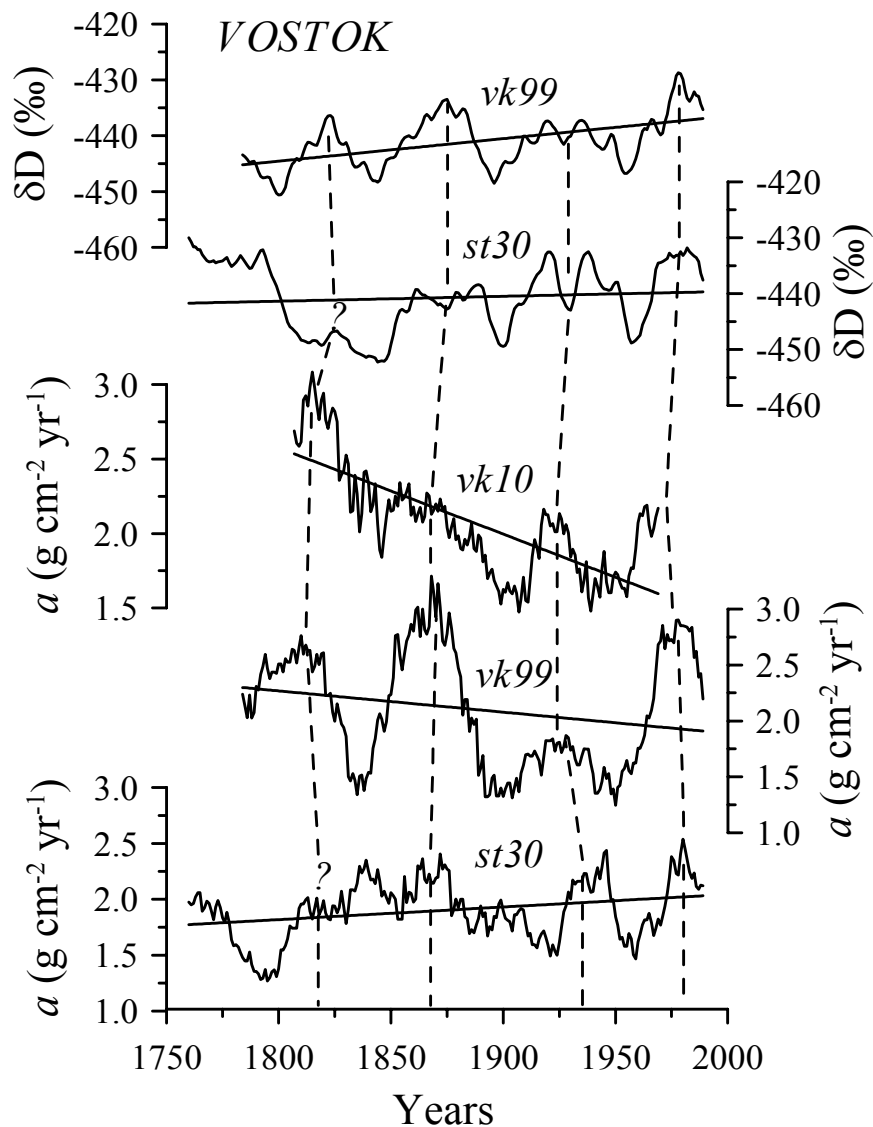


Fig. 28. Time-series of isotope composition (δD) and snow accumulation rate of snow (a) from pits $vk10$, $vk99$ and $st30$ smoothed with period of 21 years. Dashed lines link maximums of 50-year cycles of δD and a in different pits. Linear trends are shown by solid lines.

At Vostok, the influence of smaller relief forms (micro-relief and meso-dunes) on the formation of relief-related temporal oscillations of accumulation rate and snow composition of snow was experimentally proven (Chapter IV).

We thus propose that the observed secular trends of δD and a at Vostok (Fig. 25b) are due to the remaining influence of mega-dunes still preserved in the records after constructing the stacked series. The opposite sign of the trends is not in conflict with this assumption since spatial variations of δD and a are negatively correlated (Section IV.3). Our guess is also confirmed by the fact that a long-term (more than a hundred years) relief-related temporal oscillation of accumulation is predicted by the analysis of spatial distribution of snow build-up at Vostok (Section IV.2).

One could argue that the relief of ice sheet in the area of Vostok influences the long-term annual average accumulation rate itself. Indeed, the mean present-day a value at Vostok ($2.1 \text{ g cm}^{-2} \text{ year}^{-1}$) is noticeably lower than in other Antarctic sites located at similar altitudes and characterized by similar mean air temperatures. As an example, mean accumulation rates at Komsomolskaya (3498 m, -52.6°C), Dome B (3650 m, -57.5°C) and Dome Fuji (3810 m, -58°C) are, correspondingly, 6.4 [Lipenkov et al., 1998], 3.8 [Jouzel et al., 1995] and $3.2 \text{ g cm}^{-2} \text{ year}^{-1}$ [Kawamura, 2000]. At site B37 located only 111 km to NNW from Vostok the mean a value for the period 1955–1985 is $4.0 \text{ g cm}^{-2} \text{ year}^{-1}$ [Lipenkov et al., 1998]. We assume that the Vostok accumulation anomaly is caused by ice ridge to the east from the station where the glacier's sole meets the steep wall of Lake Vostok valley [Mayer, Siegert, 2000]. Since the wind in this area blows predominantly from west to east, snow is likely transported from the upwind slope of the ice ridge (where Vostok is situated) to its downwind slope.

In order to diminish the influence of relief and to establish climatic change of snow accumulation rate over the last 200 years, we collected accumulation data from 6 shallow boreholes (*vk12*, *vk15*, *BH-1*, *BH-2*, *BH-4* and *BH-8*), drilled in 1985–1996, where the Tambora layer was determined. The mean a value from 8 sites around Vostok (these 6 boreholes and 3 deep pits *vk10*, *vk99* and *st30*; values from *vk12* and *vk15* located very close one to another were taken as one value) for the period from 1816 to present is $2.06 \pm 0.03 \text{ g cm}^{-2} \text{ year}^{-1}$. This value is lower than $2.15 \pm 0.05 \text{ g cm}^{-2} \text{ year}^{-1}$ obtained for the period since 1955 to present from beta-activity measurements in 9 sites around Vostok, which suggests a slight increase of snow accumulation rate during the last two centuries.

VI.4. 200-year accumulation and isotope tendencies at other East Antarctic sites

For the purpose of comparing this result with the tendency of a broader geographical scale, we reviewed the data on a and δD changes over the past 200 years from nine East Antarctic sites: Plateau [Koerner, 1971], Dome C [Benoist et al., 1982; Petit et al., 1982], South Pole [Mosley-Thompson et al., 1999; Jouzel et al., 1983], Dronning Maud Land [Oerter et al., 2000], Mizuho [Watanabe et al., 1978], KM200 [Ekaykin et al., 2000], Talos Dome [Stenni et al., 2002], Law Dome [Morgan et al., 1991] and site DT001 in Princess Elisabeth Land [Zhang et al., 2003]. It is significant that for most of these sites we could not distinguish between climatic trends and those related to the drift of mega-dunes. Only three sites were found where it was possible to draw conclusion about veritable climatic tendency of a and δD changes:

At the South Pole the stacked a series from three boreholes shows considerable increase during the last 200 years (up to 20 %). Recent part of this accumulation increase is also confirmed by stake measurements.

At Dronning Maud Land stacked series of both a and δD obtained from 9 sites located over the area of few thousand km² reveal decreasing values during the XIXth century followed by a rise during the last 100 years. Interestingly, from 9 points where accumulation was studied in 4 the 200-year trends were positive, in 1 it was negative, and in the remaining 4 sites the trends were insignificant. This fact clearly points out at considerable influence of relief at accumulation changes in a given point in the time scale of interest.

At Law Dome the influence of relief is expected to be considerably reduced because of very high snow accumulation rates, low wind speed and due to the fact that small dimensions of the dome prevent the formation of large snow dunes. Indeed, two a series obtained from cores separated by distance of 18 km possess small amount of stratigraphic noise and correlate well with each other. The both series expose accelerated accumulation rate in the end of the last century comparing to the preceding period.

On the whole, the data from all the three latter sites testify to growing snow accumulation rate during the last 200 years over the most part of East Antarctica.

VI.5. Conclusion of Chapter V

Absolute magnitude of the variations of 11-year means of snow accumulation rate and surface air temperature over the last 200 years was, correspondingly, 1.5 g cm⁻² year⁻¹ (between

1.5 and 3 g cm⁻² year⁻¹) and 1°C (between -55 and -56 °C). No recent trends which could be related to the modern global warming are observed.

In series of both T and a quasi-periodical variations were found with a period of ~50 years. The relationship between these variations and Pacific Decadal Oscillation index implies a teleconnection between central East Antarctica and tropical Pacific. Our results suggest a sharp change in atmospheric circulation supplying Vostok area with moisture in the middle 1960s.

Interpretation of the secular trends of the δD and a is not straightforward due to contribution of long-term noise probably related to the influence of mega-dunes.

Review of data from other areas of East Antarctica shows that in most cases it is not possible to separate influence of relief and that of climate. In those cases when this separation is feasible, slight increasing of snow accumulation rate is observed over the last 200 years.

CONCLUSION AND PERSPECTIVES

As a result of the present study the following main conclusions have been achieved:

1. In central Antarctica precipitation are presented by two main types: precipitation from clear sky ("ice crystals", or "diamond dust") and precipitation from clouds with relative contribution of, correspondingly, $\frac{3}{4}$ and $\frac{1}{4}$ of total precipitation amount. The difference between mean annual amount of precipitation (26 mm) and accumulation rate (23 mm) is due to the summer snow evaporation. The mean annual condensation temperature in Vostok area is -39 ± 1.6 °C and thus equal to the inversion top temperature within its uncertainty. The mean annual air temperature is not significantly biased by precipitation-weighted temperature and thus is well representative in terms of isotope composition of snow.

2. Temporal oscillations of isotope composition and snow accumulation rate with periods of 2.5, 5 and about 20 years observed in a single point at Vostok area are mainly linked to the drift of spatial snow accumulation waves of different scales (meso-dunes). Climatic variations in δD time-series account for about 17 % of the total variance (corresponding value for accumulation rate is 10 %), most of power being concentrated in 10-year and 40–50-year periodicities.

3. The seasonal variations of precipitation isotope composition closely follow intra-annual changes of air temperature. The annual cycle of deuterium excess is opposite to those of temperature and δD and is likely related to changes of Vostok moisture source conditions. The summer minimum of d is also probably deepened by post-depositional processes in the upper snow thickness. The apparent $\delta D/T_s$ and $\delta D/T_i$ slopes are equal to, correspondingly, 2.12 ± 0.35 and 6.2 ± 1.1 ‰ °C⁻¹.

4. Inter-annual changes of snow isotope composition are well correlated with the surface air temperature variations with a slope of 17 to 28 ‰ °C⁻¹ depending on the smoothing period. This correlation is mainly determined by multi-decadal variations, while at shorter-term time-scale (about 10 years) isotope composition of snow at Vostok seems to be primarily governed by source rather than local conditions.

5. The absolute magnitude of mean multi-year values of snow accumulation rate and surface air temperature over the last 200 years did not exceed $1.5 \text{ g cm}^{-2} \text{ year}^{-1}$ (between 1.5 and $3 \text{ g cm}^{-2} \text{ year}^{-1}$) and 1–2 °C (between -55 and -56 °C). No recent trend that could be related to the modern global warming is observed.

6. In series of both T and a quasi-periodical variations were found with a period of ~50 years. The observed relationship between these variations and Pacific Decadal Oscillation index

implies a teleconnection between central East Antarctica and tropical Pacific. Our results suggest a sharp change in atmospheric circulation supplying Vostok area with moisture in the middle 1960s. The interpretation of the secular trends of δD and a is not straightforward due to contribution of long-term noise probably related to the influence of mega-dunes.

The main result of our study, in terms of isotope-temperature calibration, is displaying statistically significant linear relationships between the snow isotope composition at Vostok and air temperature at different levels, both at the seasonal and inter-annual scale. On the other hand, the obtained slopes are, in general, not consistent with those given by simple isotope models. We believe that this discrepancy is due to some drawbacks in tuning of those models. Several suggestions of refining the "classical" isotope method have been made in this work, which will be a subject of future studies. Thus, this thesis is a step towards bringing together different approaches of the paleo-temperature interpretation of the deep ice core isotope data.

Further investigations of the subjects outlined in the present study will be developed in the following directions:

1. Deeper analysis of the meteorological regime of the Vostok area taking advantage of the available dataset of meteorological and balloon-sounding data;
2. Study of post-depositional effects on snow isotope composition;
3. Modeling of isotope composition and amount of precipitation in central Antarctica using simple isotope models, regional climate models and GCMs;
4. Involving broader range of climatological information available for Southern Hemisphere, such as sea surface temperature, sea ice extent, circulation indexes, etc.

The obtained results will be used for interpretation of isotope data from the ice cores obtained by deep drilling at Vostok Station.

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ANNEXES

ANNEX A. BASIC NOTATIONS AND ABBREVIATIONS

AARI – Arctic and Antarctic Research Institute (St. Petersburg, Russia)

a – snow accumulation rate, $\text{g cm}^{-2} \text{ year}^{-1}$

b – annual snow build-up (cm)

C – coefficient of regression (slope) between changes of snow isotope composition (δ) and surface air temperature (T_S). C equals to the product of C_i and C_T

C_C – coefficient of regression between changes of condensation temperature (T_C) and surface air temperature (T_S)

C_i – coefficient of regression between changes of air temperature at the upper boundary of the surface inversion layer (T_i) and surface air temperature (T_S)

C_T – coefficient of regression (slope) between changes of snow isotope composition (δ) and air temperature at the upper boundary of the surface inversion layer (T_i)

d_{xs} or d – deuterium excess parameter ($d = \delta D - 8\delta^{18}O$ [Dansgaard, 1964])

GCMs – Global Circulation Models

r – correlation coefficient. In our study, we chose the confidence level for the correlation coefficient to be 95 %

LGGE – Laboratoire de Glaciologie et Géophysique de l'Environnement (Grenoble, France)

LGM – Last Glacial Maximum

LSCE – Laboratoire des Sciences du Climat et de l'Environnement (Saclay, France)

NBI – Niels Bohr Institute, Copenhagen University (Denmark)

PDO – Pacific Decadal Oscillation

RAE – Russian Antarctic Expedition

SST – Sea Surface Temperature

T_C – temperature of condensation

T_i – air temperature at the upper boundary of the surface inversion layer

T_S – surface air temperature (measured at the height of 2 m above the snow surface)

Greek characters

α – coefficient of isotope fractionation

δ (δD or $\delta^{18}O$) – isotope composition of a sample, representing the ratio of heavy isotope concentration (mole fraction) in sample to its concentration in standard water (in per mil):

$$\delta = (R_{SA} - R_{ST})/R_{ST} \times 1000, \quad \text{where } R = [{}^2\text{H}^1\text{H}^{16}\text{O}]/[{}^1\text{H}_2^{16}\text{O}] \quad \text{or} \quad [{}^1\text{H}_2^{18}\text{O}]/[{}^1\text{H}_2^{16}\text{O}],$$
 correspondingly. The basic standard for the oxygen and hydrogen in natural waters is

SMOW – (Standard of Mean Ocean Water). In SMOW, the absolute concentration of deuterium is 155.76 ± 0.05 ppm, and of oxygen 18 is 2005.2 ± 0.45 ppm.

σ - standard deviation

ANNEX B: LIST OF MEASURED ISOTOPE SAMPLES

Nomenclature of the data in the table: name of sample series (for example, Pit *st61*), then samples measured/total amount of samples in the series (for example, 1–157/157), then random error of measurements in per mil (for example, (0.8)).

Saclay = Laboratoire des Sciences du Climat et de l'Environnement (LSCE), Saclay, France

Copenhagen = Niels Bohr Institute of Copenhagen University, Denmark

Laboratory and period of measurements	Deuterium	Oxygen 18
Saclay, August–September 1999	Pit <i>st61</i> 1–157/157 (0.8) Pit <i>st73</i> 1–156/156 (0.8) Pit <i>vk31</i> 1–45/45 (0.8) Pit <i>vk33</i> 1–76/76 (0.8) Pit <i>vk34</i> 1–27/60 (0.8) Precip. RAE44 1–8/8 (0.8)	Pit <i>st61</i> 1–44/157 (0.055) Pit <i>st73</i> 1–156/156 (0.055) Precip. RAE44 1–8/8 (0.8)
Saclay, October–December 2000	Pit <i>vk99</i> 1–392/392 (0.6–0.8) Pit <i>st30</i> 1–332/332 (0.55–0.75) Pit <i>vk34</i> 28–60/60 (1.1) * Traverse Sheremetiev45 1–28/28 (0.26) Precip. RAE45 1–33/33 (≤ 1.1) Polygon10cm 1–40/80 (0.7) * Polygon10cm 41–80/80 (0.8) * Polygon34cm 1–80/80 (?) * Traverse Mirny–Vostok, Popov96/97, 1–28/28 (0.9) * Traverse King Boduin–South Pole, Alain Hubert97/98, 1–50/50 (0.5) * Traverse Terra Nova Bay – Dome C, 97/98, 1–21/21 (1.2) * Borehole 200KM 1–200/200 (?)	Pit <i>vk99</i> 1–53, 291–392/392 (0.04–0.06) * Traverse Sheremetiev45 1–28/28 (0.05) Precip. RAE45 1–27/33** (0.05) Polygon10cm 1–40/80 (0.04)
Copenhagen, August 2002		Pit <i>vk99</i> 1–392/392 (0.05) Pit <i>st30</i> 1–332/332 (0.05) * Traverse Sheremetiev45 1–28/28 (0.04) * Traverse Sheremetiev46 1–22/22 (0.05) * Traverse Popkov47 1–14/14 (0.05) * Traverse Sheremetiev47 1–11/11 (0.06) * Mini-pits RAE47 1–48/48 (0.06)

Annex B. List of measured isotope samples

		Precip. 2000 1–87/87 (0.05) *Polygon34cm 61–78/80 (0.04) *Series "Sigfus": - BH2 1–233/233 (0.05-0.075); - 4G1 1–192/192 (0.05-0.075); - 5G 1–186/186 (0.05-0.075); - 5G1 1–236/236 (0.05-0.075) *Series "Lipenkov": - 3400 m 1–193/193 (0.05); - 3492 m 1–138/138 (0.03); - 3572 m 1–163/163 (0.045); - 3621 m 1–62/62 (0.026); - 3612-2623 m 1–24/24 (0.025)
Aarhus, August 2002	Series "Sigfus": - BH2 1–233/233 (0.6); - 4G-1 1–192/192 (0.6); - 5G1 1–48/236 (0.6)	
Saclay, September–October 2002		Pit <i>st61</i> 36–89/157 (0.04) *Series "Sigfus": - 5G 17/186 (0.04) *Series "Lipenkov": - 3612–3623 m 1–18, 22-23/24 (0.04)
Saclay, June 2003	Precip. 2000 1-87/87 (1.5) *Mini-pits RAE47 1–11/48 (1.5)	Pit <i>st61</i> 90–157/157 (0.04) Pit <i>vk31</i> 1–38/45 (0.05) Pit <i>vk33</i> 1–11/76 (0.05) *Traverse King Boduin–South Pole, Alain Hubert97/98, 1–2, 5– 50/50 (0.07) *Traverse Terra Nova Bay – Dome C, 97/98, 1–21/21 (0.07) *Series Anisimov 1–9/9 (?)

* - not used in the present study

** - 6 samples were not measured because there was not enough material for analysis

ANNEX C: SYSTEMATIC ERROR OF ISOTOPE MEASUREMENTS

The systematic error of the isotope measurements was estimated as the offset either between the isotope values measured in the same laboratory during different time periods or between values measured in different laboratories.

Offset between isotope values measured in Saclay (LSCE)

DEUTERIUM

1) Values of Saclay'2000¹ are shifted relative to Saclay'1999 by 0.34 ± 0.14^2 ‰ (based on 34 re-measured samples).

2) Saclay'2003 – Saclay'1999: 1.27 ± 0.44 ‰ (11 samples).

3) Saclay'2003 – Saclay'2000: 0.14 ± 0.48 ‰ (9 samples). Thus, combining pp. 1) and 3) we can calculate that Saclay'2003 is offset against Saclay'1999 by 0.48 ± 0.50 ‰ (compare to p. 2)).

The best estimation of Saclay'2003 offset relative to Saclay'1999 is 0.875 ± 0.67 (calculated assuming equal weights of the items).

Thus, isotope values during period from August 1999 to June 2003 showed nearly linear increasing by about 0.23‰ per year (see the figure below).

OXYGEN 18

1) Saclay'2000 – Saclay'1999: 0.02 ± 0.011 ‰ (20 samples).

2) Saclay'2002 – Saclay'1999: 0.057 ± 0.006 ‰ (6 samples).

3) Saclay'2003 – Saclay'1999: 0.01 ± 0.0275 ‰ (4 samples).

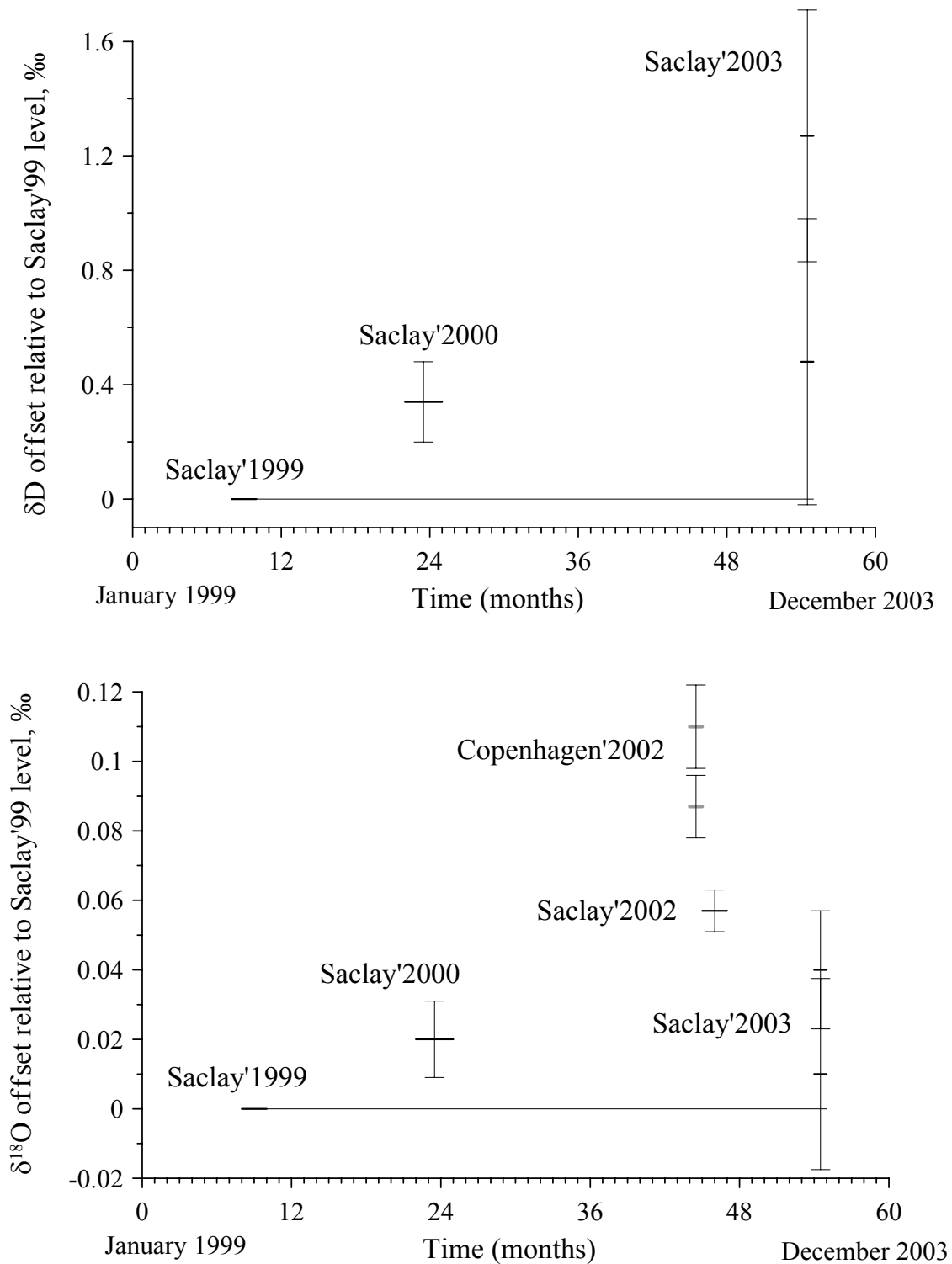
4) Saclay'2003 – Saclay'2002: -0.02 ± 0.016 ‰ (7 samples). Combining pp. 2) and 4) we obtain another estimation for the offset of Saclay'2003 relative to Saclay'1999, equal to 0.04 ± 0.017 ‰ (compare to p. 3)).

The best estimation of Saclay 2003–1999 shift is 0.025 ± 0.032 ‰ (calculated assuming equal weights of the items).

¹ Saclay = Laboratoire des Science du Climat et de l'Environnement (LSCE), Saclay, France; Copenhagen = Niels Bohr Institute of Copenhagen University, Denmark. See Annex A for the list of samples measured in each laboratory.

² Uncertainty of mean offset value is calculated as $\frac{\sigma(\Delta\delta)}{\sqrt{n}}$, where $\sigma(\Delta\delta)$ is standard deviation of the offset values for individual samples, n – number of samples

Thus, $\delta^{18}\text{O}$ values during the period from August 1999 to October 2002 increased linearly by about 0.018 ‰ per year, and then decreased between October 2002 and June 2003 by roughly 0.03 ‰ (see the figure below).



DEUTERIUM EXCESS

From the observed offsets of the both isotopes it is possible to calculate the resulting offset in deuterium excess values. During the period from August 1999 to October 2002 deuterium excess

must have shifted only slightly (by 0.09 ‰ per year) because ratio of δD offset relative to the $\delta^{18}O$ one was not very different from 8 ($0.23/0.018 = 13$). In October 2002 total shift of the measured deuterium excess values relative to the August 1999 level could reach about 0.26 ‰. For the measurements made in June 2003, deuterium excess values may be offset by about 0.7 ‰ (from 0.16 to 1.2 ‰ if taking different estimations for δD and $\delta^{18}O$ offsets by June 2003). Sharp increase of deuterium excess shift in June 2003 is explained by the fact that between October 2002 and June 2003 shift of $\delta^{18}O$ became less while that of δD kept increasing.

The observed offsets cannot be explained by changing of isotope composition of standard water with time, because storing would lead to increasing of isotope composition of standard water (both due to evaporation and isotopic exchange with the ambient air) and, consequently, to lower measured values of the isotope composition of the samples. Probably, the shifts are caused by storing of the samples themselves. In this case, it is difficult to explain the strange behavior of the $\delta^{18}O$ offset between October 2002 and June 2003. Also, we should note that all the samples are stored in frozen state in special hermetic plastic containers, which reduces to the minimum the possibility of sample water evaporation or exchange. If the observed shifts are indeed related to the storing effect of the samples, then all the shifts mentioned above represent change of isotope composition of samples as a function of storing time, and do not concern fresh samples.

Offset of oxygen isotope values measured in Saclay (LSCE) and Copenhagen (Niels Bohr Institute)

1) Copenhagen'2002 is offset by 0.03 ± 0.007 ‰ relative to Saclay'2002 (37 samples).

2) Copenhagen'2002 – Saclay'2000: 0.09 ± 0.005 ‰ (182 samples).

Recalculating these values relatively to the level of Saclay'1999 gives us the following estimation of Copenhagen'2002–Saclay'1999 offset: 0.09 ± 0.009 ‰ and 0.11 ± 0.01 ‰. The best estimation is thus equal to 0.10 ± 0.015 ‰. This shift is hardly can be explained only by storing of the samples and suggests systematic error in the standard water value used in one of (or the both) the laboratories. If the above assumption about the origin of the Saclay isotope value offsets as related to the time of sample storing is correct, then the real difference between isotope values produced by the two laboratories is 0.03–0.05 ‰, $\delta^{18}O$ values from Copenhagen being larger than those from Saclay.

Implications for our study

As a result, we have not introduced any corrections in the isotope values used in the present study provided that a given series of samples was measured in one laboratory during comparatively short period of time. In case if a series was measured during several periods of

Annex C. Systematic error of isotope measurements

time separated by comparatively long intervals (for example, samples from pit *st61*), results of earlier measurements were reduced to the level of later measurements (to do this, we ensured that later measurements partly overlap earlier ones). In case of *vk99* pit samples and precipitation'2000 samples, they were partly measured in Saclay and then completely measured in Copenhagen; thus, we reduced values obtained in the first laboratory to those measured in the latter. This possibly can produce a bias of oxygen 18 values from pits *vk99* and *st30* (also measured in Copenhagen) and precipitation'2000 of up to 0.05 ‰ relative to the other series measured in LSCE (corresponding offset in deuterium excess values would be up to $8 \times 0.05 = 0.4$ ‰).

In general, we conclude that the maximum estimation of systematic error (due to sample storing and errors in standard water values) for the isotope data discussed in the present study is about ± 1.3 ‰ for δD values and ± 0.11 ‰ for $\delta^{18}O$ values. Estimated systematic error for the deuterium excess values does not exceed ± 1.2 ‰. These figures are (considerably) less than the natural variability of snow isotope composition at Vostok and thus can be recognized as satisfactory for our study.