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SCIENTIFIC AREA: **Earth Sciences (Global Change)**

PROJECT TITLE:

PALEO-CLIMATIC and ENVIRONMENTAL ICE CORE RESEARCH, DATA ANALYSIS and INTERPRETATION

(mid- low- latitudes, high altitudes glaciers at the Northern Hemisphere)

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REPORT

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Table of Contents:

1. Preface

One of the most valuable information about past climate and environmental changes lies in glacial ice, which records preserve precipitated snow for hundreds to thousands of years. These records from pre-industrial or even prehistoric time can be examined through snow-ice stratigraphy and geochemical analysis for stable and radioactive isotopes, major ions, trace elements and green house gases (GHG). We can link this information with changes in atmospheric circulation, air temperature, snow accumulation, atmospheric composition, marine and continental biogenic activity, aerosol loading/volcanic eruptions, continental dust source regions, forest fire activity, anthropogenic emissions, solar variability, radionuclide deposition and the GHG chemical composition of the atmosphere. Supplementary long-term (50 to 150 years) information including meteorological, hydrological, and atmospheric chemistry observational data is used for our statistical calibration, validation and interpretation analyses.

Our project is a multi-disciplinary, multi-institutional, international effort in ice-coring paleo-climatic environmental research and we working to help better understand the impacts of global and environmental changes on the natural ecosystems. In this report, we present some results of our research performed in FY2002/2003.

Our research goal *is to recover ice-core isotope-geochemical records containing information on large-scale atmospheric dynamics, the precipitation-origin, the natural and anthropogenic impact on climatic variability during industrial and pre-industrial time to understand past and forecast future Global Changes*

GLACIO-MONITORING 2001, 2002 and 2003 DEEP ICE-CORING EXPEDITION in SIBERIAN ALTAI

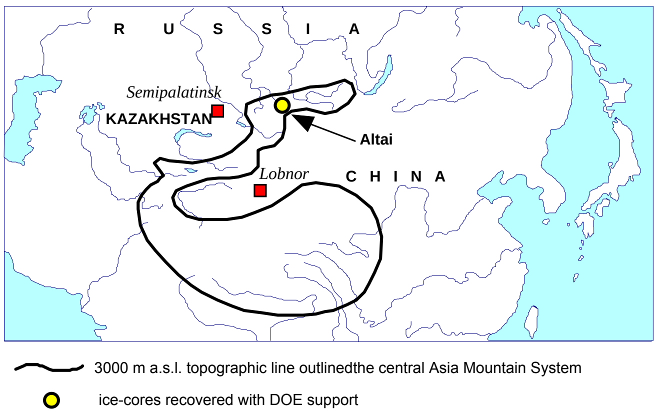

Figure 1. Location of the Asian research sites

Overview

After successfully recovering two deep ice-cores from central Tien Shan in the summer 2000 we focused our research on Siberian Altai and south-eastern Tibet-Himalaya (Figure 1).

The spatial coverage of available snow, firn and ice records is inadequate to document climatic and environmental change over the vast Asian continent. However, environmental records for the northwestern periphery of the central Asia mountain system (CAMS), obtained from Tien Shan firn/ice cores (*Kreutz and others*, 2001, 2003; *Aizen and others*, in press a, b) and firn/ice-core records from alpine areas in Siberia (*Aizen and others*, in press b; *Olivier and others*, 2003), are extending the area of climatic and environmental analyses in Asia. The research presented in this chapter focuses on the Siberian Altai, the most continental northern periphery of the CAMS and the southern periphery of the Asian Arctic Basin (Figure 1). It is an ideal area for the analysis of climatic records relating to the major Eurasian circulation systems (i.e. the westerly jet stream and the Siberian High). Altai glaciers, located at the center of Eurasian continent, store unique information on inter-hemispheric climate dynamics and on the internal and external hydrological cycles of northern Eurasia. They provide records on the advection of fresh water transported from the Atlantic, Pacific and Arctic Oceans. Firn-ice records from the Altai glaciers can also be associated directly with the large Aral–Caspian internal water system, because moisture is transferred

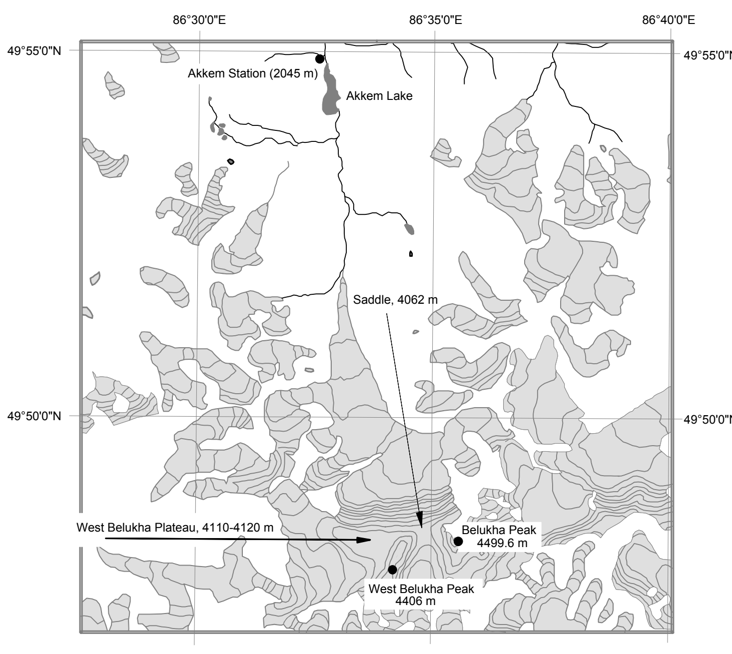

Figure 2. West Belukha Plateau location, Siberian Altai

from this closed Asian drainage basin to the great Siberian river basins. Furthermore, the Siberian Altai is only mountain system in the Asian Arctic Basin (Figure 1) with alpine glaciers that nourish the Siberian rivers. Altai glaciers, particularly their cold snow-firn accumulation zone are appropriate for studying air pollution dynamics at the center of Eurasia, eastward and northward from major air pollutants in Russia, Kazakhstan and China. During the 20th century the Altai Mountains became extremely contaminated from industrial activities including mining, metallurgy and chemical production, as well as from the nuclear test areas in Semipalatinsk (Kazakhstan), Lobnor (China), and the Baikonur rocket site (Russia).

The major components reflecting physical processes in the atmosphere over the studied region are $\delta^{18}O$ and $\delta^{2}H$ stable-isotopes, major ions, and trace elements transported and deposited on the glaciers**.** Data on the stable-isotope and geochemical content in snow and ice cores have contributed to our understanding of environmental signals from different regions of the globe (*Aizen and others*, 1996, in press a, b; *Dansgaard and others*, 1993; *Froehlich and others*, 2002; *Johnsen*, 1977; *Jouzel and others*, 1997; *Naftz and others*, 2002). Information on the spatial distribution of stable isotopes and geochemical components in precipitation in the Siberian Altai is also required to evaluate the contribution of water vapor, associated with external and internal water cycles, to snow accumulation over the northern periphery of the CAMS.

Field research 2001, 2002

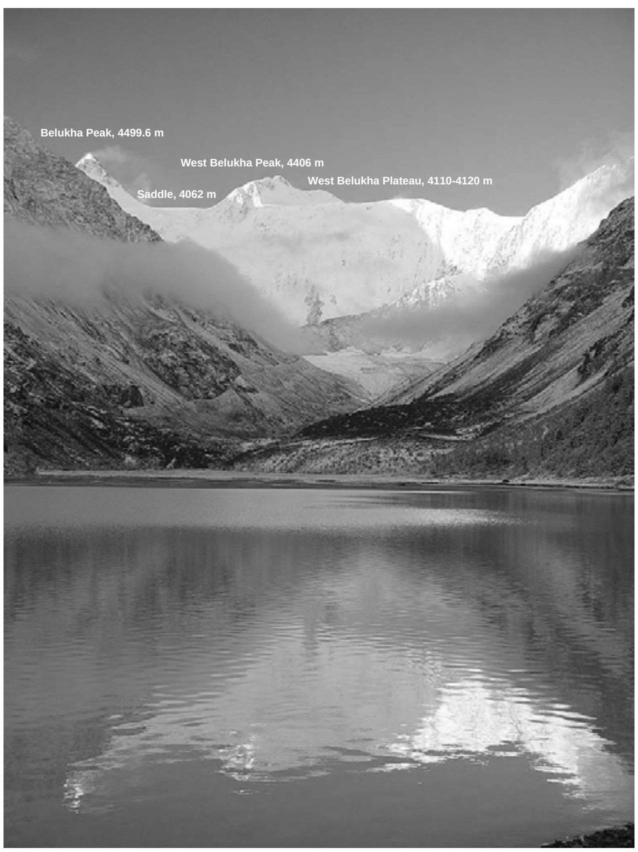

Figure 3. Belukha glacial massif, Siberian Altai. View from Akkem Lake (northern side). Photo by V. Aizen, August 2003.

The West Belukha plateau **(**Figures 2 and 3) is the only Siberian location where the Altai glaciers have cold enough temperatures and sufficient snow accumulation to preserve climatic and environmental records that are unaffected by meltwater percolation. The field site selected by *Oliver and others* (2003), on a saddle (4062 m a.s.l.) between the Belukha (4499 m a.s.l.) and West Belukha (4406 m a.s.l.) peaks (Figures 2, 3 and 4), was a little too low and, because of very strong snow and wind redistribution between the two peaks, unsuitable for ice-core research.

In the summers 2001, 2002 reconnaissance and 2003 deep ice-coring expedition, snow samples were collected every 3–5 cm from five 2–3 m snow pits that dug in the same location each year on the Belukha snow/firn plateau (4110–4120 m a.s.l.). In 2001, a 21 m shallow snow/firn core was recovered at 4115 m a.s.l. (49°48’ N, 86°33’E), where radio-echo sounding indicated the ice thickness is about 180 m thick (Figure 4). Snow pit, precipitation, and fresh snow samples were collected and placed into pre-cleaned plastic bottles. Density was measures every 5 cm in each snow pits using a 100 cm^3 stainless steel sampler. A firn core (19 m long and 9.5 cm in diameter) was extracted from the bottom of the 220 cm snow pit (number 4) with a PICO fiberglass auger. The diameter, length and weight of each recovered core section were measured to calculate density. The 2001 firn-core sections, sealed in pre-cleaned polyethylene bags, were packed in insulated shipping containers and delivered to the National Institute of Polar Research in Tokyo (NIPR) for analysis (*at that time we didn't have facilities in UofI for this work*).

A Grant Instruments Co. automatic weather station was installed at 4100 m a.s.l., near a rocky cliff on the northern side of Plateau, about 0.5 km north of the drilling site (Figure 1.4 and 1.5). Wind speed, wind direction, air temperature, air humidity and barometric pressure sensors measured meteorological events every 3 hours from July 2002 to August 2003 (Figure 5 and 6).

Temperatures in the snow pits were measured every 10 cm from surface to bottom using electric sensors. The West Belukha snow/firn plateau is in the cold recrystalization zone where melt does not occur; positive temperatures were never observed at the drill site (Figure 6).

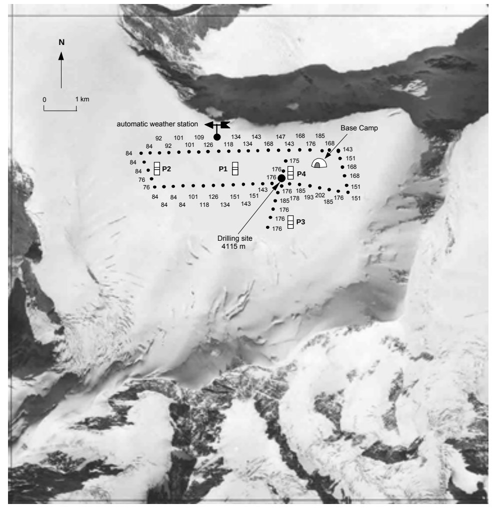

Figure 4. Glacier ice thickness profiles, snow pits and ice-core sites, on the Western Belukha Plateau.

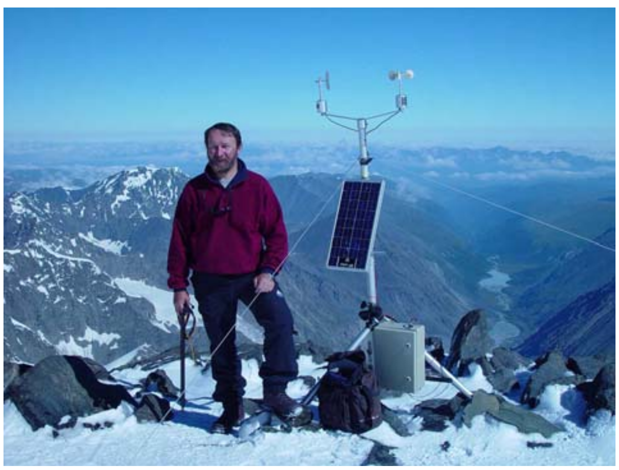

Figure 5. Automatic weather station installed at 4100 m a.s.l., 0.5 km north from the drilling site. July 2002, photo by V. Aizen

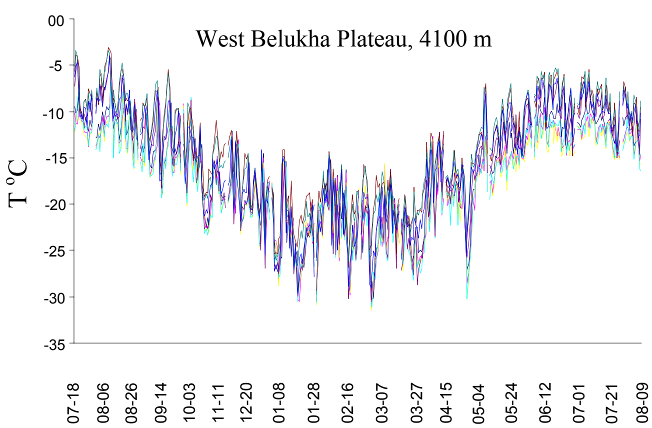

Figure 6., Hourly air temperature. July 2002 to August 2003.

To monitor snow accumulation, we installed 5 wooden stakes, with location determined by GPS (Garmin eTrex summit). One stick was equipped with an automatic snow depth measurement KADEC-SNOW, KONA system. The geo-coordinates of these stakes, snow pits and drill site location are summarized in Table 1. The geodesic coordinate accuracy was ±0.01 m in horizontally directions and ±0.05 m in vertically.

The accumulation gauge sensor detects snow or open air by photo-diodes at an interval of 1 cm. Since strong solar radiation during noontime infiltrates into snow causing an apperent lowering of the surface level, the noontime records were discarded. Two records, morning and evening were chosen and validated to represent the actual surface level. Unfortunately, accumulation gauge sensors stopped on October 18, 2001, recording only 3 months records but even this data shows that precipitated snow can be drifted out from the plateau by strong winds (Figure 7).

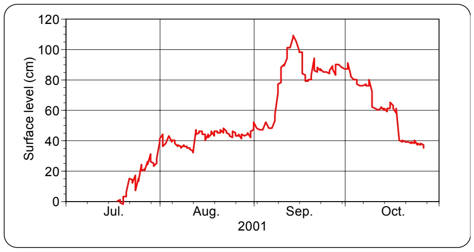

Figure 7. Three month accumulation records at 4150 m a.s.l. of the Plateau.

Long-term meteorological data

For climatic analysis, we used long-term (50 years) meteorological data from the Akkem (49°54' N, 86°32' E; 2045 m a.s.l.), Kara–Turek (49°57' N, 86°29' E; 2600 m a.s.l.) and Aktru (°50'5 N, 87°46' E; 2110 m a.s.l.) stations located 10, 15 and 30 km north, northwest and northeast of the West Belukha snow/firn plateau. Data from these stations has the highest correlation with air temperature and precipitation in the Altai glacierized area (*Galakhov*, 1981) particularly with the West Belukha Plateau weather station. All long-term meteorological data was checked for homogeneity, inspected for the presence of random errors, and plotted for periods with missing observations, following recommendations by *Easterling and others* (1995). The air temperature gradient (T_{ds}) at the drill site (H_{ds}) was calculated from Equation (1)

$$T_{ds} = T_0 - \gamma(t) \cdot (H_{ds} - H_0), \qquad (1)$$

where T_0 (°C) is mean air temperature at the altitude of Akkem station (H_0); γ(T), 0.0068°, 0.0072°, 0.0039° and 0.0021°C m^{-1}, are the spring, summer, autumn and winter mean altitudinal gradients of air temperature, calculated using long-term data from the Akkem, Kara–Turek, and Aktru meteorological stations, and one year of measurements by the West Belukha Plateau weather station.

To describe atmospheric circulation patterns over southwestern Siberia that might influence regional precipitation regimes at seasonal time scales, we used monthly data of the atmospheric pressure distribution from the North Atlantic Oscillation (NAO), East Atlantic Pattern (EA), East Atlantic/West Russia Pattern (EA/WR) and Pacific–North American (PNA) indices. The principal component scores of the patterns were obtained from http://www.cpc.noaa.gov/data/teledoc/telecontents.html, which were derived from rotated empirical orthogonal function analysis (*Richman*, 1986).

2003 ice-coring operation

In the summer of 2003, after two years reconnaissance and glacio-climatic monitoring in Siberian Altai, Russia two 175 m surface to bottom ice cores have been successfully recovered from the Western Belukha Plateau at elevation 4150 m a.s.l. An electro-mechanical drill with 9.5 cm diameter (inner size) and 135 cm long barrel manufactured by Geo Tecs Co. (Japan) was used for this operation. The

Table 1. Geodetic locations by measurement in 2001 (Datum: WGS84).

Location	deg N	min N	deg E	min E	Altitude	Date
Gorno-Altaisk Airport	51	58.246	85	50.293	339	11-Jul
Akkem Base Camp	49	54.361	86	32.518	2041	11-Jul
Karatjurek Pass	49	56.037	86	30.334	3068	13-Jul
Glacier Camp	49	48.331	86	33.722	4087	16-Jul
Pit 1	49	48.251	86	33.667	4107	19-Jul
Pit 2	49	48.258	86	33.536	4109	19-Jul
Pit 3	49	48.330	86	33.506	4113	19-Jul
Pit 4	49	48.333	86	33.652	4114	19-Jul
Pit 5	49	48.308	86	33.694	4115	19-Jul
Stake 1	49	48.286	86	33.694	4110	19-Jul
Stake 2	49	48.253	86	33.663	4101	18-Jul
Stake 3	49	48.214	86	33.632	4102	19-Jul
Stake 4	49	48.263	86	33.609	4109	19-Jul
Stake 5	49	48.242	86	33.717	4112	19-Jul

maximum core length for one run with this drill is approximately 55 cm. It took 87.5 hours actual working time (7 working days) to drill one core. The total number of the drill runs was 325. Core sections totaled 324, with a mean length of 48.6 cm. Most core sections were not brittle and were recovered in perfect physical condition.

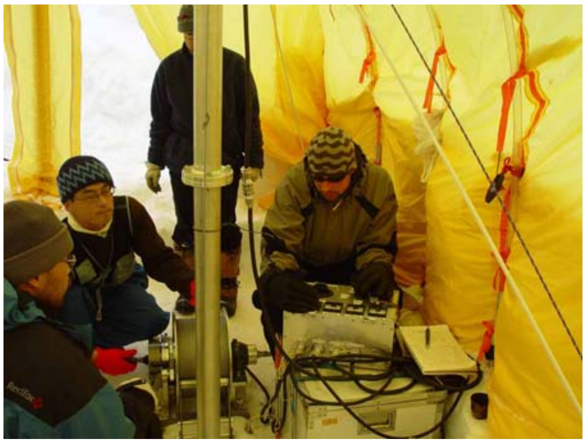

Figure 8. Drilling process. Daniel Joswiak, Arzhan Surazakov, Tetsuhide Yamasaki and Akiyoshi Takahashi. West Belukha Plateau. Photo by V. Aizen

The 2003 expedition started on July 20 from Novosibirsk, a large industrial city in south-western Siberia. From Novosibirsk all expedition gear and personnel were transferred to Gorno Altaisk, the capital of Altai Republic in the Russian Federation. A bus was rented to transport personnel about 700 km south from Novosibirsk to Gorno Altaisk. The expedition gear and food were transported with a large commercial truck. After one night in Gorno Altaisk we flew to Akkem base camp in a Russian MI-MTV commercial helicopter. Two shuttles between Gorno-Altaisk and Akkem base camp, near the Western Belukha Plateau were required to carry out all personnel (14 people) and about 5 tons of expedition gear, food and fuel. July 21-24 was spent in Akkem cabins preparing for operation on the Plateau. Two days of preparation at Akkem base camp, located at elevation 2,045 m a.s.l. (Figure 2), we spent sorting our cargo and hiking for acclimatization. Weather was rainy and foggy, which is quite usual for this time of year in the Altai Mountains. Nevertheless, on July 25 the helicopter came and we all moved to the Plateau in several flights. It was the early morning, and a strong blizzard blowing up snow required us to move slowly and carefully windward, dragging all our supplies to the location where we had our basecamp in two previous expeditions. All day until sunset we worked hard to establish our high elevation camp at 4,150 m a.s.l. Walls made from snow blocks were set up around each tent to prevent them from being buried in drifting snow. A snow trench measuring 2x3x5 m was dug and covered by half inch plywood to store food and kitchen supplies. Two ‘NothFace’ 5m dome tents were putt side-by-side for cooking and dining facilities. A 2.5 kW electric generator was used for lighting and drilling day and night. The next day, July 26 a large, semi-cylindrical drilling tent was set up, measuring 4.5 x 3.5 m x 2.25 m. To secure space for the drill system, the snow surface level was dug out and lowered by 70 cm. This tent is designed to withstand winds in excess of 20 m sec^{-1}, and was additionally reinforced with lumbers and ropes in case of even stronger winds which can occur on the plateau. Another snow trench 2x2x5 m was dug in front of the drill shelter to function as a field snow laboratory, where each ice core section was weighted, described, and photographed for preliminary stratigraphic analysis (Figure 8). Finally, core sections were packed in thermo-insolated boxes to awaite transportation of the plateau. Plywood boards were used as a roof of the trench. To maintain a clean environment, the drill ten and snow laboratory were placed 200 m south of our base camp and electric generator.

Drilling started on July 28 (Figure 8). At least four people were required for the drilling operation. One person is required to operate the controls while three people maneuver and handle the drill body, which is laid horizontally after each run The drilling operation was usually done from 6 am to 12 pm and from 4 pm to 12 am.

The first ice core was drilled to a depth of 174.3 m in depth on August 4. At this depth, the cutter began to chip from hitting basal rock debris. After obtaining this core, the borehole temperature was measured every 10 m. Drilling for the second core started on August 5. The second core was drilled approximately 4 m east of the first borehole, and finished on the August 10 at the same depth.

All ice core sections were stored in 30 insulated boxes (size: L 129 x W 50 x H 50, Insulation Shipping Containers Co., U.S.A). The boxes were transported by a helicopter from the plateau to Gorno-Altaisk Airport. From there they were carried by a commercial freezer truck to Novosibrisk International Airport, where they were stored in an airport freezer at –20 degree until transportation to Japan and USA by air cargo (Figure 9).

Figure 9. Cargo terminal, Seattle International Airport. Ice-core arrived. Photo by V. Aizen

Ice core properties

Most of the cores sections are not brittle and were obtained in a good cylindrical shape (Figure 10). The cores from surface to about 50 m are mixed of ice and firn layers. Dipper than 50 m, the firn becomes bubbly ice with some clear ice layers.

Figure 10. Dust layer in core section 111 at 51.4 m depth.

Figure 11. Alexei Chebotarev with ice-core.West Belukha Plateau. Photo by V. Aizen

Bulk density of the cores increased with depth, reaching 0.9 g/cm^{-3} at approximately 50 m. The density increased with depth in a roughly linearly fashion from 7 m to 42 m.

Four visible dust layers were observed in the ice core, excluding dust and dirty layers at bottom part the core (Figure 10 and 12). The depths of the dust layers were 51.4, 54.6, 57.8, and 122.4 m. The ice core drilled at the eastern

Figure 12. Loosed stones close to the bottom at 169.6 m depth.

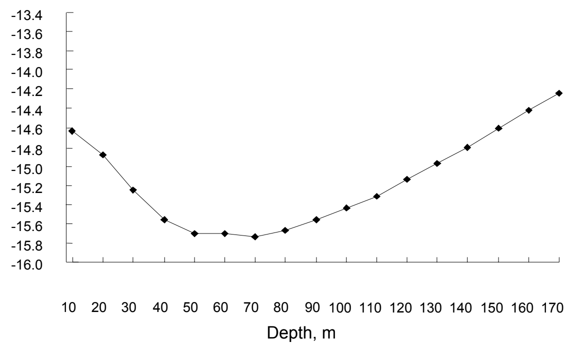

Figure 13. The borehole temperature profile. West Belukha Plateau

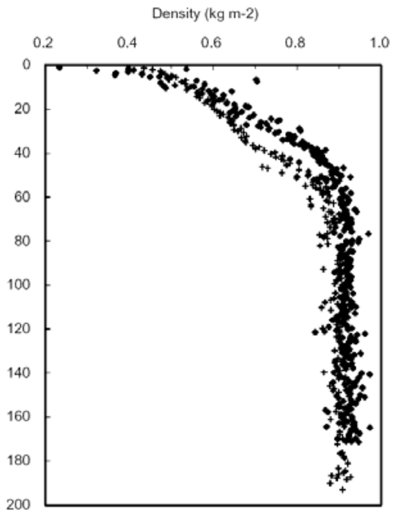

Figure 14. Density of deep ice-core. West Belikha Plateau.

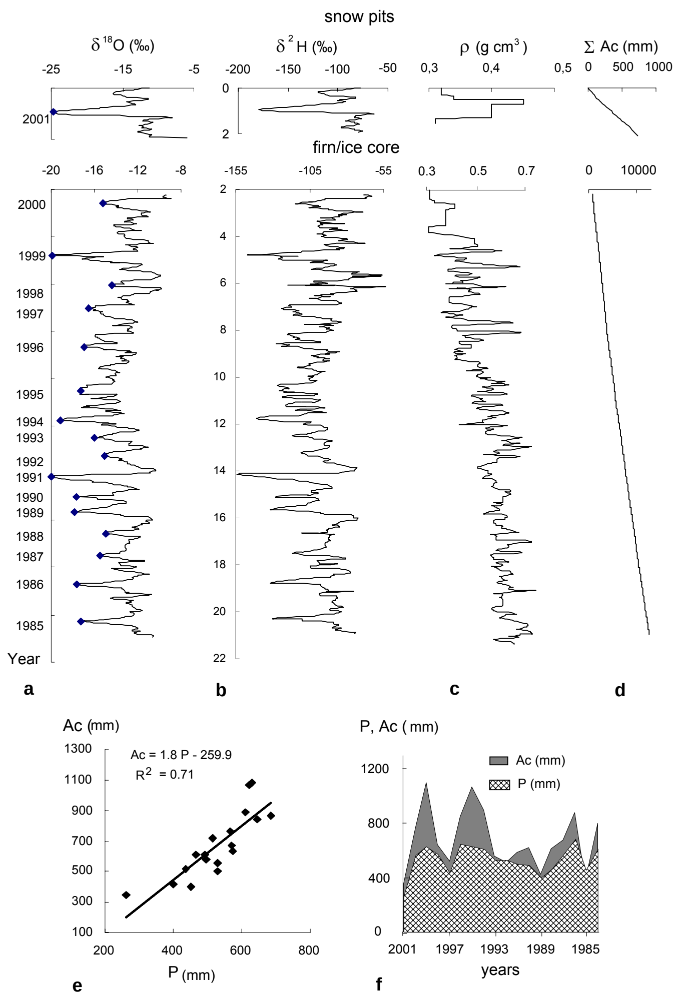

Figure 15. Isotopic composition, $\delta^{18}O$ (a); $\delta^{2}H$ (b); snow/firn density, (c); cumulative snow water equivalent, ΣAc (d), in the snow/firn core. (e) Correlation between annual accumulation, Ac, records from firn/ice core with precipitation, P, at the Akkem station. (f) Distribution of annual accumulation records from firn/ice core and precipitation at the Akkem station. Years (1984–2001) represent winter intervals.

plateau of Mt. Belukha by *Olivier et al.* (2003), also contained dust layers, which are probably corresponded with the dust layers of our ice cores. These dust layers will be further analyzed discussed, and interpreted in the near future

Core sections deeper than 163 m from the surface were visibly dirty, containing silt and sand, suggesting that this ice is near the basal ice/rock interface. The core sections from 169.6 m depth contained small stones about 1 cm in diameter (Figure 12).

Borehole temperatures were measured every 10 m, the change with depth is depicted in Figure 13. Temperatures decreased with depth to 70 m, reaching the minimum of –15.7 °C at 70 m, then increased with depth to -14.2 °C.

The laboratory analysis and first results

Stable isotope analysis

The Altai firn core 2001 and snow samples of 2001 and 2002 were processed in a dedicated cold room at NIPR using techniques established for ultra-clean sample preparation. Frozen 18 mΩ water blanks were passed through the entire system to ensure there was no contamination and for quality control. The melt index (*Koerner and Fisher*, 1990), compiled from the measured 5 cm density (Figures 14 and 15) and analyzed stratigraphy (Figure 16) showed < 5% melt. Each 3–5 cm of the upper 11 m of the snow/firn core, as well as samples from five snow pits, fresh snow and precipitation, were analyzed for $\delta^{18}O$ and δ^2H at the NIPR. 10 m of the bottom section of the core were analyzed for δ^2H at the University of Maine and for $\delta^{18}O$ at the University of Idaho. The technique of snow/firn-core stable-isotope analysis has been described by *Kreutz and others* (2001).

In the clean room facilities at the Geochemical Laboratory of the National Polar Research Institute in Tokyo each snow and ice sample processed for the major ions and geochemical elements was brought to dryness (using acid-cleaned PTFE vessels) via sub-boiling evaporation inside a HEPA-filtered laminar flow box. Acid digestion of sample particulates was performed by adding a 1:3 HF:HNO3 solution (Optima brand) and heating (_~55^{o}C) overnight. Thus, major and trace element data from these samples represent total sample concentration (soluble plus insoluble). Although we have not quantified the digestion efficiency of this technique, visual inspection suggests that any remaining material was largely organic. Samples were then redissolved in 1N HNO3 (Optima and Milli-Q) for analysis with a Finnigan Element high-resolution magnetic sector inductively coupled plasma mass spectrometer (ICP-MS). The rare earth element suite was measured in low-resolution mode ($m/\Delta m = 300$), while Ca, S, Al, and Fe were measured in medium resolution ($m/\Delta m = 3000$). Detection limits (given as sample blank 3σ) for each element are given in Table 1; total sample blank values were <1% for all elements measured in each sample. The sulfur isotope composition of sample filtrate was determined after filtering through 0.4-mm pore size PTFE filters. Two contrasting firn layers were analyzed: one from a visible dust layer (7.61±7.80 m WED), and one from a section of core with relatively low SO_4^{2-} concentrations representing ``background'' dust conditions (6.87±7.39 m WED). Each sample represents the combination of two adjacent sections of the core. The total amount of S in the samples (measured as SO_4^{2-} via ion chromatography) was 190.8 and 15.3 mg, respectively. Both samples were evaporated to a volume of _2 mL, and transferred into tin reaction vessels. Samples were pulse combusted in an elemental analyzer, and SO_2 gas was separated chromatographically. The SO_2 was analyzed via gas-source mass spectrometry and is reported in delta notation versus the Canyon Diablo Troilite (CDT) standard. Estimated ratio error for the two samples is ±0.05%.

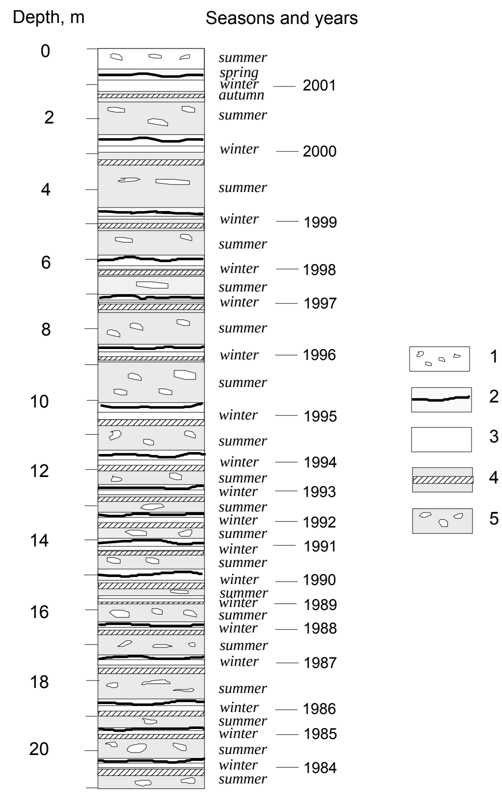

Figure 16. Stratigraphy of 21 m snow/firn core recovered from the West Belukha plateau with seasonal and annual layer identification: 1 is snow with ice lenses up to 5 mm thick; 2 is fine-grained firn with 1–2 mm ice crusts; 3 is fine-grained compact white firn; 4 is fine-grained firn with aeolian particles; 5 is coarse-grained firn with ice lenses up to 8 mm thick.

Establishing a depth–age relationship

No significant post-depositional effects were apparent in the 2001 firn-core oxygen and hydrogen records, as the amplitude of the signal near the bottom of the core was similar to that observed near the surface (Figure 15a and b). To determine annual snow accumulation layers, we used stable-isotope records with existing, well-preserved seasonal δ^{18}O and δ^{2}H signals (Figure 15a and 15b), visible seasonal accumulation evidence (Figure 16) and annual precipitation data from the Akkem station (Figure 17b). Identification of annual accumulation layers in the snow/firn core was based on the lowest δ^{18}O and δ^{2}H values. Minimum winter air temperatures are a distinctive characteristic of the Altai meteorological regime (Figur 17a). Snow and firn–ice densities (Figure 15c) were used to establish cumulative depth/water-equivalent profiles (Figure 15d). A significant correlation of 0.71, between stable-isotope values and the snow/firn density, allowed the identified annual layers to be verified. Maximum densities, observed during warm seasons, are associated with high isotopic values, while minimum densities, related to cold seasons, are associated with low isotopic values. Snow/firn stratigraphy was also used to identify annual layers in the snow/firn core (Figure 16). The highest correlation, between annual accumulation layers in the snow/firn core and precipitation at the Akkem station, reached 0.82 (Figure 15e and 15f). The mean annual accumulation rate, calculated at the drill site, was 690 mm from summer 1984 to summer 2001. To determine the seasonal and monthly oxygen–deuterium distribution in the snow/firn core, a relationship between snow accumulation on the Belukha plateau and precipitation at the Akkem station was developed. The mean accumulation rate obtained agrees with the calculated accumulation rates presented in the *World Atlas of Snow and Ice Resources* (Kotlyakov, 1997).

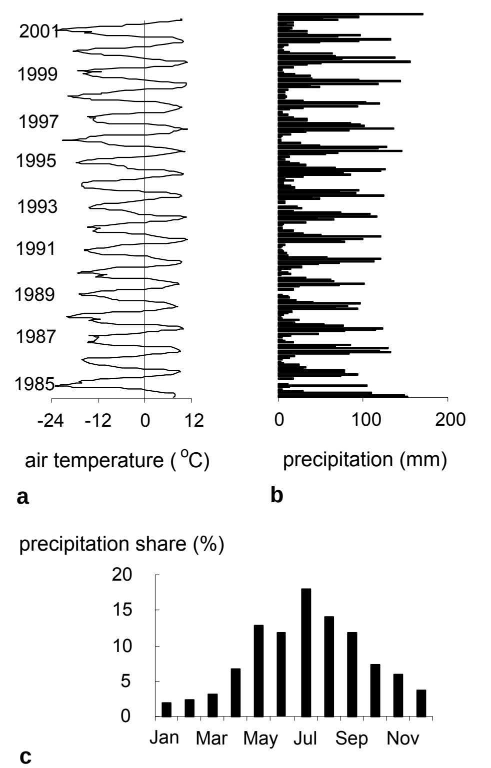

Figure 17. Monthly mean air temperature (a), precipitation (b), and monthly percentages of annual precipitation (c) at Akkem station.

Seasonal snow accumulation

The monthly and seasonal snow accumulation values from summer 1984 to summer 2001 at 4115 m a.s.l. were calculated using the linear relationship obtained from the snow/firn core and precipitation at the Akkem station (Figure 15e), and then normalized according to *Aizen and others* (in press a). Calculated seasonal snow accumulation values were verified using the snow/firn stratigraphic profile (Figure 16), where winter layers could be differentiated from others by their homogenous bright-white crystal structure. In the spring layers, there are signs of 1–2 mm slender radiation crusts, while summer layers have ice lenses up to 8 mm thick. The autumn layers could be identified by more compact firn and yellow-brown aeolian particles. Uncertainty in calculating seasonal accumulation using the above relationship (Figure 15e**)** was less than ±10% of the seasonal accumulation rate obtained from the snow/firn core stratigraphic profile. The corresponding annual seasonal $\delta^{18}O_{season}$ and $\delta^{2}H_{season}$ compositions were averaged as arithmetic means of every 3–5 cm measured stable-isotope ratio in each seasonal accumulation layer of the snow/firn core. The same approaches have been used by Aizen and others (in press a), *Barlow and others* (1993), *Shuman and others (1995) and Yao and others* (1999) to evaluate seasonal accumulation in snow/firn/ice cores.

Table 2. Oxygen and deuterium isotope ratios ($\delta^{18}O$, ‰) in snow pits, firn/ice cores, fresh and old snow, and precipitation from the Belukha plateau, Altai and other glaciers along the northern periphery of central Asia and northern Tibet

	Snow pit 1	*Snow pit 2*	*Snow pit 3*	*Snow pit 4*	*Ice core*	*Fresh snow*	*Old snow*	*Precipi-tation*
			Altai, Belukha plateau, 4115 m a.s.l. ($\delta^{18}O$, ‰)					
n	40	40	40	40	484	3	6	28
Ave	-15.5	-14.6	-15.3	-14.0	-13.6	-15.3	-23.7	-15.3
Max.	-8.8	-6.3	-8.6	-8.1	-9.1	-14.9	-22.4	-6.1
Min.	-27.9	-25.9	-25.4	-24.7	-19.9	-16.6	-25.3	-18.3
St dev.	4.8	4.8	3.7	3.9	1.9		1.3	3.9
Δ	19.2	19.6	16.8	16.6	10.8	1.3	2.9	12.2
			Altai, Belukha plateau, 4115 m a.s.l. ($\delta^{2}H$, ‰)					
n	40	40	40	40	484	3	6	28
Ave	-112.1	-105.8	-110.4	-101.7	-97.9	-101.9	-176.2	-114.3
Max.	-59.9	-48.9	-61.9	-63.5	-55.1	-101.9	-161.7	-53.7
Min.	-204.5	-194.3	-191.3	-179.6	-146.7	-115.6	-194.6	-140.8
St dev.	35.9	36.4	28.8	29.1	15.6		11.5	25.7
Δ	144.6	145.4	129.4	116.1	91.6	13.7	32.9	87.1
			Altai ($\delta^{18}O$, ‰) (Kotlyakov and Gordienko, 1982)					
Tomich glacier, 2300–2750 m a.s.l.			-16.4 -26.1					
			Altai ($\delta^{18}O$, ‰) (Oliver, 2003)					
Belukha saddle, 4062 m a.s.l.					-12.0			
			Tien Shan, Inylchek glacier ($\delta^{18}O$, ‰) (Kreutz,and others,2001; Aizen and others, in press b)					
n	40		20	20	205	25		
Ave	-13.05		-14.18	-12.67	-16.8	-15.24		
Max.	-7.96		-9.58	-10.11	-6.0	-13.92		
Min.	-18.41		-17.74	-14.91	-35.6	-16.00		
St dev.	2.24		2.50	1.35	6.7	0.57		
Δ	10.45		8.16	4.79	29.6	2.08		
			Pamir ($\delta^{18}O$, ‰) (Kotlyakov and Gordienko, 1982)					
Abramova glacier, 4400 m a.s.l.				-10.5 -16.8				
Glacier, head Lyadjuardara river, 5200 m a.s.l.				-13.3 -27.4				
Krasnoslobodceva glacier, 5040 m a.s.l.				-10.0 -25.5				
Akbaital glacier, 5100 m a.s.l.				-18.0 -21.0				
Bakchigir glacier, 5000 m a.s.l.				-16.4 -26.1				
			North Tibet ($\delta^{18}O$, ‰) (Thompson and others,1995)					
Guliya ice cap					-13.1			
Dunde ice cap					-12.5			

Note: *n* is number of samples. Ave, max., min., st dev., Δ are average, maximum, minimum, standard deviation and amplitude.

Stable isotopes from firn cores and snow pits

The mean oxygen-isotope ratio from the Altai core of –13.6‰ is in accordance with oxygen-isotope records obtained from central Asia and the northern Tibetan mountains (Table 2). The higher average annual air temperature at the Altai drill site, at a lower altitude than the Tien Shan drill site (*Aizen and others*, in press a), corresponds to a slightly higher mean oxygen-isotope ratio than the –16.8‰ measured in the Inilchek glacier core, in the Tien Shan. For the Altai firn core, the highest $\delta^{18}O$ and $\delta^{2}H$ seasonal accumulation variability was observed in winter firn layers (Table 3a). The annual variation in $\delta^{18}O$ from the snow pit and firn core (19.6 and 10.8‰) was about the same as that from the Pamir and other Altai glaciers (Table 2). The Tien Shan glaciers exhibit a greater variation in $\delta^{18}O$ (29.6‰) than other Asian sites (Table 2) which is attributed to the lowest winter isotopic means.

Table 3. Belukha plateau snow/firn core monthly average $\delta^{18}O$ and $\delta^{2}H$ composition and their linear relationship

	Jan	*Feb*	*Mar*	*Apr*	*May*	*Jun*	*Jul*	*Aug*	*Sept*	*Oct*	*Nov*	*Dec*
***(a)** Monthly average $\delta^{18}O$ and $\delta^{2}H$ composition*												
						$\delta^{18}O$						
Ave	-16.9	-16.8	-16.7	-15.5	-13.8	-12.9	-13.1	-13.1	-13.9	-14.2	-15.9	-16.3
Max.	-14	-14	-14	-12.7	-10.8	-9.1	-10.3	-10.1	-9.8	-8.1	-10	-10
Min.	-24.7	-22.8	-21	-19.9	-17.4	-17	-17.9	-19.1	-20.1	-19.5	-19.9	-23.6
St dev.	2.6	2.3	1.9	1.7	1.3	1.9	1.5	1.6	2.1	2.6	2.4	2.9
						$\delta^{2}H$						
Ave	-122.7	-122	-120.5	-110.8	-99.6	-92.1	-94.3	-95	-101.1	-102.3	-113.1	-115.7
Max.	-93.5	-93.5	-93.5	-88.6	-74.4	-55.4	-71.6	-72	-68.9	-55.1	-55.1	-55.1
Min.	-179.6	-168	-151.4	-149.9	-128.5	-123.6	-134	-139.7	-149.3	-153.9	-149.9	-176.5
St dev.	21.3	19.5	16.5	14.7	10.9	15.3	12.1	12.8	16.8	21.9	22.2	26.2
***(b)** Linear relationship of $\delta^{18}O$ and $\delta^{2}H$ composition*												
n	17	17	18	31	56	105	101	107	49	32	17	16
a	7.8	8	8	8	7.8	7.9	7.8	7.9	8.3	8.1	8.1	8.3
b	8.6	12.7	13.3	13.7	8.6	10.9	7.9	8.8	7.8	13	13.3	18.1
R^2	0.92	0.91	0.88	0.87	0.91	0.94	0.93	0.92	0.97	0.94	0.9	0.95

Note: *n* is number of measurements, R^2 are coefficient of determination.

Higher winter minimum $\delta^{18}O$ and $\delta^{2}H$ values in the Altai core (–19.9‰) than in the Tien Shan cores (–35.6‰), caused by higher winter air temperatures at the Altai site and more short trajectories of air masses bringing precipitation there, demonstrate the continental effect of meteoric water depletion moving farther from the source of water vapor (*Dansgaard,* 1964; *Friedman and others,* 1964). During the cold season, air masses are moving eastwards across the Eurasian continent along the main route: high latitudes from Iceland to western Siberia and then to central Asia (*Glukh and Kononova*, 1982). Air masses with water vapor precipitated over the Tien Shan have therefore been subject to a stronger depletion of $\delta^{18}O$.

The highest seasonal mean $\delta^{18}O$ and $\delta^{2}H$ in the Altai (Table 3a) and Tien Shan (*Aizen and others*, 2004a) snow/firn cores were for the summer accumulation layers. However during the warm season, the air masses are crossing the Eurasian continent eastward along two main routes: high latitudes from Iceland to western Siberia and mid- (even low-) latitudes to middle Asia (*Glukh and Kononova,* 1982). Altai glaciers receive moisture from higher latitudes than do the Tien Shan glaciers, where water vapor is transferred from lower latitudes and altitudes. Hence, the maximum $\delta^{18}O$ values in the Altai cores are slightly lower than those observed in the summer layers of the Tien Shan core (Table 2).

Oxygen- and hydrogen-isotope ratios from Altai snow pits (Figure 8) varied significantly; the lowest values (Figure 15 top) corresponding to the lowest minimum air temperature and an abnormally high amount of winter precipitation observed in winter 2000/01 (Figure 17). The extremely low hydrogen- and oxygen-isotope ratios (–28 and –200‰) in winter 2000/01 were caused by low air temperatures and by a significant amount of direct precipitation not enriched by evaporation.

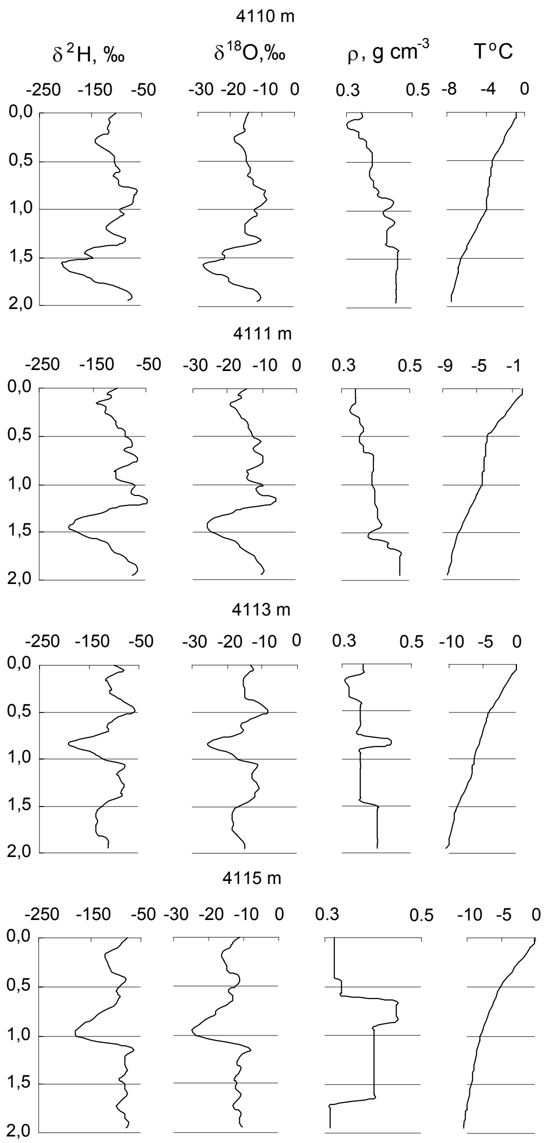

Figure 18. Deuterium (δ^2H), oxygen ($\delta^{18}O$), snow/firn density (ρ and snow temperature T (°C)) measurements in the West Belukha plateau snow pits, 2001.

Relationship between stable hydrogen- and oxygen-isotope ratios

The relationship in the precipitation, snow pits and snow/firn core from the Belukha plateau (Figure 19) has the same slope to the co-variance (i.e. 8) as that of the global meteoric water line (GMWL, Equation (2)) described by *Craig* (1961):

$$\delta^2H = 8.0\ \delta^{18}O + 10 \qquad (2)$$

Similar slopes observed in local relationships (Figure 20), and in the GMWL, indicate the same initial relationship of fraction factors and point to the absence of strong melt and percolation in the snow/firn core and snow pits. The δ^2H–$\delta^{18}O$ relationships related to June accumulation layers in the snow/firn core records (Table 3b**)** are closest to the GMWL. The smaller or larger intercept in the local relationship, from the snow/firn core (Table 3b), fresh snow, snow pit and precipitation, than in the GMWL () reflects different kinetic evaporation effects on the water vapor transferred to the Altai mountains, e.g. initial water vapor was quickly or slowly evaporated under non-equilibrium conditions (*Kendall and McDonnell*, 1998).

Clustering precipitation transferred from external (oceanic) and internal (continental) moisture sources

The equations corresponding to evaporation from the ocean under conditions close to equilibrium were obtained by *Dansgaard and Tauber* (1969). They were based on precipitation data from 38 island and continental stations around the North Atlantic (Equations (3 and 4)):

$$\delta^2H = 8.1\ (\pm 0.1)\ \delta^{18}O + 11\ (\pm 1) \qquad (3)$$
$$\delta^{18}O = 0.69\ T - 13.6 \qquad (4)$$

To distinguish between oceanic moisture and water vapor evaporated from internal drainage basins (e.g. local basins or the Aral–Caspian basin) and transferred to the northeastern latitudes of Asia, the δ^2H–$\delta^{18}O$ isotopic relationships from the Altai snow/firn core were clustered into three distinct datasets (Figure 20a). The clustering procedure was based on inequalities (5–7) resulting from Equation (3).

$$\text{“Atlantic”:}\ \ 8.2\ \delta^{18}O + 10 \le \delta^2H \le 8.0\ \delta^{18}O + 12 \qquad (5)$$

"Oceanic": $\delta^2H < 8.2\ \delta^{18}O + 10$ (6)
"Internal": $\delta^2H > 8.0\ \delta^{18}O + 12$ (7)

where $\delta^{18}O$ and δ^2H are the isotopic contents in 254 samples from the snow/firn core.

The deuterium excess (d, ‰) was also defined by $d = \delta^2H - 8\delta^{18}O$ (*Dansgaard,* 1964). Precipitation from the Altai snow/firn-core records, having a hydrogen- and oxygen-isotope relationship based on the conditions of inequality (5) (the same as Equation (3)), was considered to have been transferred from the northern part of the Atlantic Ocean. A smaller intercept in the snow records (inequality (6)) reflects a smaller kinetic effect during evaporation, when conditions were closer to equilibrium than in the records corresponding to Equation (3). Precipitation with this record also originated from oceanic water. Precipitation with a hydrogen- and oxygen-isotope relationship based on the conditions of inequality (7) was more enriched in deuterium than in records corresponding to Equation (3). High deuterium excess indicated strong kinetic effects which were most probably caused by water evaporation from internal moisture sources, e.g. Aral and Caspian basins or convection from a local basin.

Total snow/firn accumulation in the Altai core from June 1984 to July 2001 amounted to 12,591 mm w.e., 69% corresponding to inequalities (5) and (6), i.e. precipitation transferred from external, oceanic moisture sources, and 31% to inequality (7), i.e. precipitation transferred from internal moisture sources (Fig. 9b). There is no significant dependence of deuterium excess on oxygen content (Fig. 9c).

Internal moisture source [inequality (7)]

The records associated with internal moisture sources have the highest deuterium excess (15.04‰, Figure 20b) as a result of low air humidity at the center of the continent during water vapor evaporation from local moisture sources.

The smallest share of precipitation transferred from the internal water cycle to the Altai

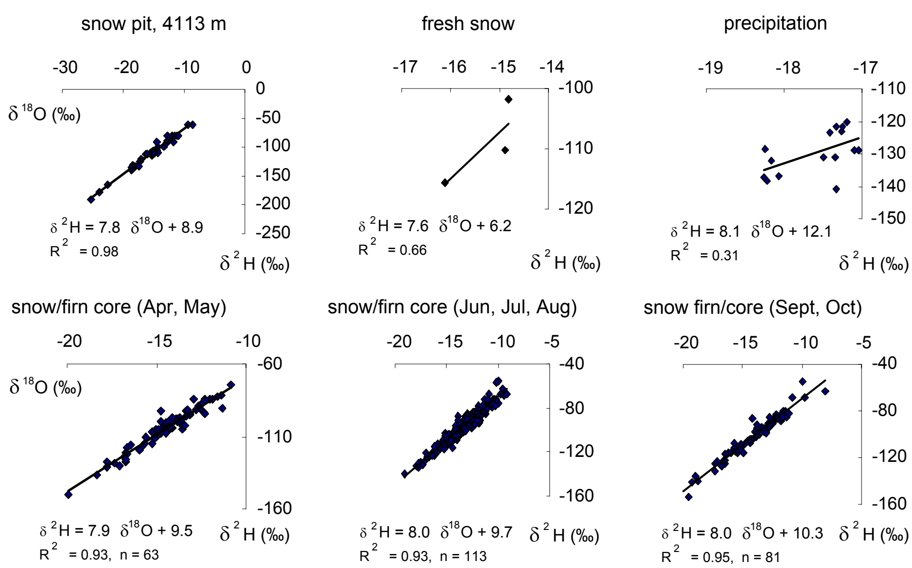

Figure 19. Stable deuterium (δ^2H) and oxygen-isotope ($\delta^{18}O$) relationship in snow pit, fresh snow, precipitation and snow/firn core on the West Belukha plateau, 2001.

glaciers was observed in winter (27%, **Figure 20b**), when the Siberian High is strongest over northern and central Asia, blocking any intrusion of air masses, and conditions for inland evaporation and local convection are weakest because of low continental heating. The mean deuterium excess is highest in winter (up to 16.7‰, **Fig. 20b**), when air humidity during

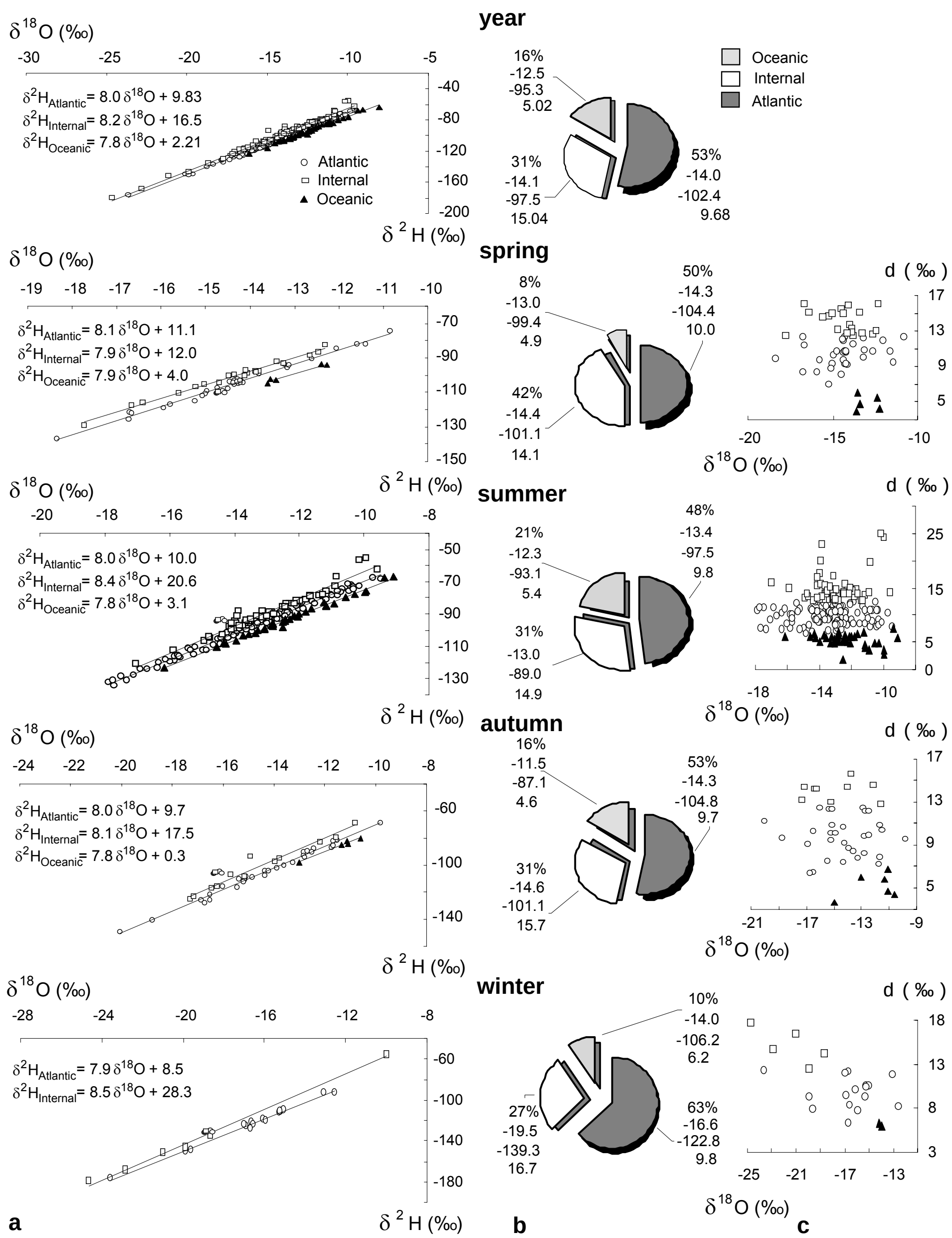

Figure 20. (a) Annual and seasonal clustered stable δ^2H and $\delta^{18}O$ relationships. (b) Annual and seasonal contributions to accumulation of precipitation originating from external and internal moisture sources: percentage of average oxygen, deuterium and deuterium excess. (c) Seasonal relationship between deuterium excess and oxygen isotopic composition (d) in snow/firn core from the West Belukha plateau, 2001.

evaporation over inland moisture sources is lowest. Absolute summer deuterium excess reaches 23‰ (Figure 20c), indicating a strong kinetic effect during evaporation, when air humidity is low over the inland moisture flow located at the center of Eurasia. This is in accordance with results from *Froehlich and others* (2002) on the maximum mean of deuterium excess obtained from stations in the Northern Hemisphere.

In spring, warming of the inner continental regions occurs faster than in the coastal regions, which increases inland evaporation. Therefore, the largest annual proportion of precipitation is brought to the Altai from internal moisture sources in spring (42%, Figure 20b). Among the seasonal records, the mean intercept in the oxygen-hydrogen relationship and deuterium excess are lowest in spring (Figure 20b).

In summer and autumn, the contribution of internal continental water vapor to accumulation on central Asian glaciers is also high (up to 31%). Absolute summer deuterium excess reaches 25‰ (Fig. 10c) indicating low air humidity over the inland moisture flow.

Atlantic moisture source [inequality (5)]

Moisture with a hydrogen- and oxygen-isotope relationship corresponding to water vapor evaporated over the northern Atlantic Ocean ($\delta^{18}O = 8\delta^{2}H +9.83$, Figure 20a) comprises more than half the annual accumulation (53%, Figure 20b). The $\delta^{18}O$–$\delta^{2}H$ relationship varied slightly during a year (Figure 20a). Deuterium excess during a year was almost invariable, ranging from 9.7–10‰ (Figure 20b). The mean values of the $\delta^{18}O$ records related to precipitation origin, from internal and Atlantic Ocean sources, are about the same during all seasons (Fig. 10b). The range of variability in their absolute values is also similar, fluctuating from –9 to –25‰ (Figure 20c). During the winter, $\delta^{18}O$ records associated with inter-land moisture sources are even lower than $\delta^{18}O$ records related to North Atlantic water vapor origin. Water vapor evaporated from the southern internal moisture sources in winter, probably moves northward over the cold Asian continent and mixes with cool air masses, accounting for stronger depletion in $\delta^{18}O$ than water vapor evaporated from the relatively warmer ocean.

Oceanic moisture source [inequality (6)]

Precipitation transferred from an external moisture source, with the smallest deuterium excess, comprises on average 16% of the annual accumulation with a summer maximum (up to 21%, Figure 20b) and spring minimum (8%). The lowest mean d-excess (4.6‰) was observed in autumn records; most probably associated with the highest atmospheric moisture saturation over the ocean during water vapor evaporation, further moisture transportation and precipitation over the Altai. The mean and absolute values of $\delta^{18}O$, from accumulation layers (Figure 20b and c) having an "oceanic" moisture origin, are the highest among the three considered clusters, ranging from –9‰ to –15‰. The highest $\delta^{18}O$ values may be associated with the highest air temperature during precipitation (*Aizen and others*, 1996; *Lin and others*, 1995; *Yao and Thompson*, 1992; *Yao and others*, 1995) and/or near-by water-vapor formation, e.g. Pacific Ocean.

Relationship between $\delta^{18}O$ and air temperature (i.e. transfer function)

Multiple picks (Figure 15) with high $\delta^{18}O$ and $\delta^{2}H$ isotopic relationships in annual variations of stable-isotope records were the result of: (1) summer maximum air temperature; (2) several sources of moisture including oceanic and inter-land, partially during spring, summer and autumn, and; (3) different trajectories by which air masses were transporting moisture to the Altai mountains.

Transfer functions and continental/oceanic moisture sources

The transfer functions (Figure 21a) were developed for three considered clusters of precipitation origin (section 6.1) using monthly data on air temperatures at the drill site (Equation (1)) and isotopic ratios averaged for seasonal accumulation snow/firn layers (section 5.2). Annual and seasonal air temperatures for each cluster were averaged from the monthly means, if the records related to a considered cluster were observed in these months. Transfer functions (Equations (8–10)) inside each cluster differed insignificantly (range of variation in parentheses) during all seasons (Figure 21a):

"Atlantic": $\delta^{18}O = 0.6\ T - 13.0\ (\pm 0.4)$ (8)

"Oceanic": $\delta^{18}O = 0.85 (\pm 0.05) T - 10.2 (\pm 0.4)$ (9)
"Internal": $\delta^{18}O = 1.05 (\pm 0.05) T - 10.0 (\pm 0.6)$. (10)

The most similar relationship, to the co-variance found in precipitation from the North Atlantic (Equation 4), was observed during all seasons in the records clustered under the condition of inequality (5) (Equation 8), pointing to North Atlantic precipitation. Under the same air temperatures, the most negative $\delta^{18}O$ values are associated with precipitation from the North Atlantic, while the heaviest values were associated with moisture transferred from an oceanic source (Fig. 11a) that could be the Pacific or Arctic Oceans. Considering that relatively heavy $\delta^{18}O$ isotopic ratios are characterized by higher temperatures, we expect this precipitation origin was the Pacific not the Arctic Ocean.

The transfer functions developed for precipitation from the inter-land moisture sources lie between the two functions developed for oceanic originated precipitation (Figure 20a).

The different isotope–temperature slopes (i.e. 0.6‰, 0.85‰ and 1.05‰ °C^{-1}) in the transfer functions (Equations (8–10)) are the result of a different intensity in the enrichment of snowfall by heavy isotopes through snow evaporation and sublimation. With "Internal" air masses (Equation (10)), low humidity, insignificant amounts of precipitation and snow/wind storms caused intensive evaporation and sublimation of falling snowflakes. "Atlantic" air masses (Equation (8)) are associated with intensive snowfalls, high humidity and relatively windless weather. With these, much of the falling snow was not enriched by evaporation and sublimation.

Transfer functions and large-scale atmospheric patterns

To verify different sources of moisture nourishing the glacier being studied, we compared mean values of $\delta^{18}O$ in snow/firn core seasonal accumulation layers with seasonal mean air

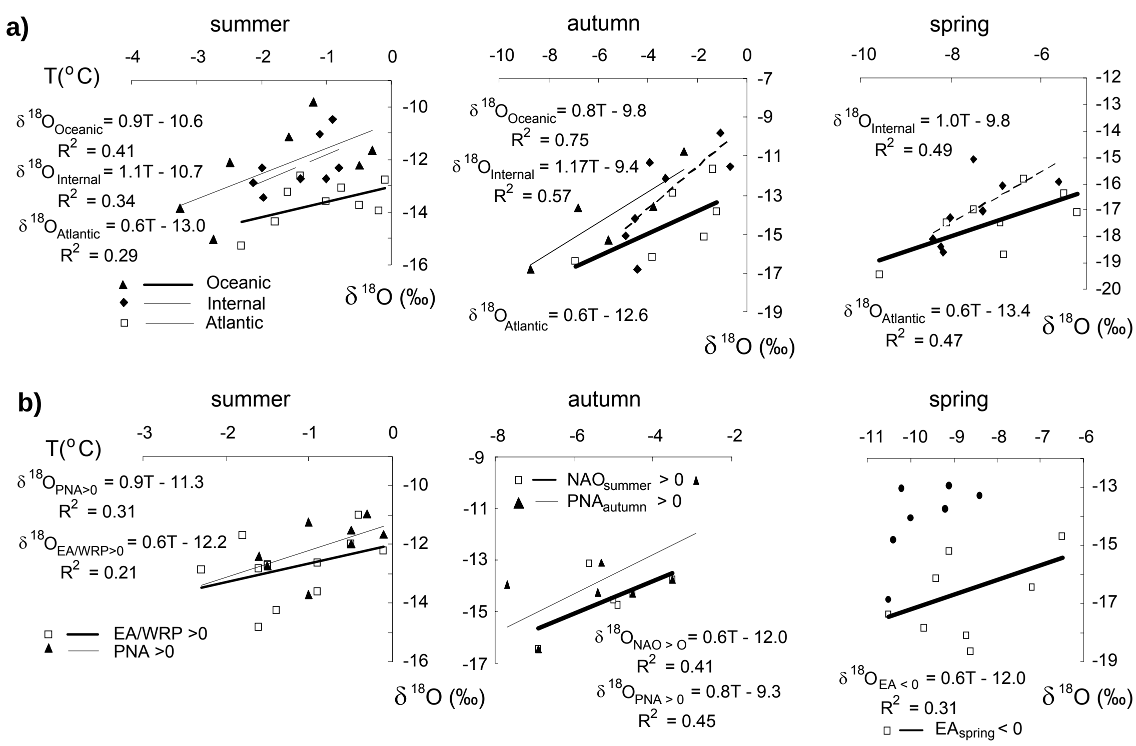

Figure 21 Seasonal relationships between $\delta^{18}O$, content in snow/firn cores and air temperature at 4115 m a.s.l. depending on (a) sources of moisture and (b) prevailing large-scale atmospheric patterns.

temperatures at the drill site (Equation (1)), considering the indices of atmospheric circulation patterns. The sign of the seasonal EA, EA/WR, NAO and PNA values differentiated these relationships (Figure 21b). The negative spring values of EA and positive summer values of the NAO and EA/WRP indices are associated with moisture brought from the west (i.e. Atlantic Ocean, Mediterranean and Black Seas) to the Siberian Altai. Positive values of PNA indices are associated with moisture brought from the Pacific.

'NAO, EA/WRP > 0; EA < 0': $\delta^{18}O = 0.6\,T - 12.1\ (\pm 0.1)$ (11)

'PNA > 0': $\delta^{18}O = 0.85\ (\pm 0.05)\,T - 10.3\ (\pm 1.0)$ (12)

The relationships based on records associated with Atlantic moisture sources (Equation (8), Figure 21a) are most similar to those that considered the NAO, EA/WRP indices (Equation (11), Figure 21b). Furthermore, linear transfer functions associated with an oceanic moisture source (Equation (9), Figure 21a) are comparable with equations relating $\delta^{18}O$ and air temperature during the seasons when the PNA indices had positive values (Equation (12), Figure 21b).

Findings

Analysis of the oxygen–deuterium content in a 21 m snow/firn core allowed us to estimate the annual/seasonal amounts and origins of precipitation nourishing the Belukha plateau region of the Siberian Altai.

Annual accumulation was found to be about 700 mm w.e. at 4115 m a.s.l. for the period from summer 1984 to summer 2001. This is in agreement with the meteorological and snow/firn stratigraphy data.

The mean oxygen isotope ratio of –13.6‰, as well as an annual $\delta^{18}O$ variation range of 19.6‰, from the Altai core is in accordance with oxygen-isotope records obtained from central Asia and the northern Tibetan mountains. The slopes in the hydrogen and oxygen relationship (i.e. 8), in the precipitation, snow pits and snow/firn core and d-excess of 10.63‰, are similar to slopes and d-excess in the GMWL co-variance. However, the clustering analysis of the oxygen–deuterium relationships indicated that both Atlantic and Pacific Oceans and inner-continental water are the sources of moisture to the Siberian Altai. The 17 year snow/firn accumulation, recorded in the 21 m core, comprised 69% of the precipitation transferred from the oceans. In the annual accumulation, more than half was from water vapor evaporated over the northern Atlantic Ocean. The contribution of precipitation transferred from the Pacific Ocean varied from 8% in spring to 21% in summer, when the lowest d-excess (4.6‰) was associated with high atmospheric moisture saturation over the ocean during water-vapor evaporation. Among the three clusters, the highest mean and absolute $\delta^{18}O$ values were related to precipitation originating from the Pacific and were associated with relatively higher air temperature during snowfall. The inter-land moisture sources contributed 31% of annual precipitation, with the spring maximum reaching 42%. Low humidity at the center of the continent resulted in the highest d-excess (up to 25‰) in the records associated with inter-land moisture sources.

The linear transfer functions, between air temperature and isotopic composition in accumulated precipitation from both external and internal moisture sources, were developed at the seasonal temporal scale. Under the same air temperatures, the lightest $\delta^{18}O$ values are associated with precipitation from the Atlantic Ocean, while the heaviest values are associated with that from the Pacific Ocean. Low humidity and snow/wind storms caused significant evaporation and sublimation of falling snowflakes and a bias on exceeding the isotope–temperature slope when air masses transferred moisture from "Internal moisture sources" rather than when air masses originated over the Atlantic. The relationships based on records associated with Atlantic moisture sources are most similar to those which considered the NAO, EA, EA/WRP indices, whereas linear relationships associated with the Pacific moisture origin are comparable with equations based on records when the PNA indices had positive values.

According to identification of annual layers in the cores, the mean annual snow accumulation for the period from 1997 to 2001 was found to be about 800 mm. Analysis of large-scale atmospheric patterns (NAO and PNA), oxygen and deuterium isotope content in ice core and snow pits assume two sources of moisture on the Altai Mountains.

Further proposed laboratory and analytical research (Altai)

1. complete deep ice-core (175 m) analysis for major ions, trace elements, stable isotopes (^{18}O and ^{2}H), Green House Gases (GHG), black carbon, radionuclide, heavy metals
2. ice-core data calibration and validation
3. data interpretation and simulation
4. technical reports and manuscripts preparation for publication in peer review journals

Necessary time: **3 years (2004-2006)**

References:

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