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A study of longitudinal and altitudinal variations in surface water stable isotopes in West Pamir, Tajikistan

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ABSTRACT

In an effort to establish a clear relation between stable isotopes and altitude in the Pamir region, as well as to improve the understanding of stable isotope spatial variations found along the routes taken by the westerlies, surface river water samples were collected along a roughly west–east profile in Tajikistan. Here we present the spatial changes in modern surface water $\delta^{18}O$ and deuterium excess (*d*-excess), and their links with snow melt, glacier melt and seasonal precipitation patterns in Tajikistan. The results show a close relation between river water $\delta^{18}O$ and δD above the GMWL (Global Meteoric Water Line), implying that surface evaporation exerts a weaker influence. Spatially, river water $\delta^{18}O$ is gradually depleted from west to east due mainly to the effect of altitude, yielding a river water $\delta^{18}O$ vertical lapse rate of 0.09‰/100 m. River water *d*-excess shows a decreasing trend from west to east in Tajikistan. This spatial *d*-excess pattern in surface water is explained by the different moistures influenced by the Mediterranean and other in-land seas with higher *d*-excess in west, and no-Mediterranean moisture in east. Another possible reason is the differences in precipitation seasonality between the west and east of Tajikistan.

1. Introduction

Many efforts have been made to understand the relation between stable isotope variations and the hydrological cycle and atmospheric circulation on the Tibetan Plateau (TP) and its surrounding area (Li et al., 2007; Warburton and DeFelice, 1986; Xiao et al., 2014; He and Keith, 2015). Previous work has mainly focused on differences in moisture transport in the monsoonal and westerly regions, and its significance in paleoclimate research (Araguas-Araguas et al., 1998; Davis et al., 2005; Tian et al., 2007; Vuille et al., 2005; Yao et al., 1996; Bershaw et al., 2012). Large spatial changes along the Indian monsoon trajectory have been identified using latitudinal precipitation and land surface water isotopes. Altitudinal changes in water isotopes are also used to constrain the uplift of the plateau in that region (Quade et al., 2007; Rowley and Currie, 2006; Rowley et al., 2001; Holdswoith et al., 1991). A clearer understanding of how isotopes vary with elevation is now required; such a variation has been well-defined on the Indian monsoon-influenced southern slopes of the Himalaya or the southern TP by sampling river water along the vapor trajectories (Garzione et al., 2000; Kent-Corson et al., 2009; Rowley et al., 2001), but there remains a lack of information in the westerly region. Stable isotopes have also been used to rebuild paleoclimate changes from ice core records at different alpine glacier sites (Henderson et al., 2006; Thompson et al., 1989, 1997; Hren et al., 2009). Although long-term ice core

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isotope records are believed to reflect temperature change (Thompson et al., 2000), spatial changes in different ice core isotope records are also apparent, as derived from different atmospheric circulation controls (Yao et al., 2006). This hints at the investigative isotope work necessary for regions with different atmospheric circulations. Spatial river water *d*-excess can also be used to identity different types of moisture transport as it can bear different isotope signals (Tian et al., 2001; Salamalikis et al., 2015).

Tajikistan is located in Central Asia, and has a typical arid and semi-arid temperate continental climate. The country's topography is typified by a sizeable contrast in altitude between the low-lying plains in the southwest and the high Pamir Mountains in the east (Peak Somoni, the highest peak, with an altitude of 7495 m asl, is in the central Pamir). This large altitude contrast makes it an ideal region for isotope vertical gradient rate estimation. A huge number of glaciers surround the high mountains of Pamir (Aizen et al., 1996). The Fedchenko Glacier, the longest alpine glacier in the world outside of the polar regions, has the potential to yield long-term paleoclimate change signals from its ice core isotope records. The meltwater from these glaciers is Tajikistan's main available water resource. In fact, Tajikistan accounts for more than half of the total available water resource in the entire Central Asia region. Two large-scale water suppliers to the Aral Sea, the Syr Darya and Amu Darya rivers, originate in the glaciers of the western Tianshan and Pamir Mountain ranges. It is thus important to understand the isotopic hydrological cycle of this region.

Here we present river water isotope ($\delta^{18}O$ and δD) data along a west–east transect in Tajikistan. The factors influencing river water isotopes, including precipitation isotopes and their seasonal changes, vapor sources and land surface evaporation, are very complex. A thorough understanding of these factors and how they determine water isotopes is thus problematic, owing especially to a lack of continuous precipitation isotope data. However, such isotope data as there are can demonstrate links with local water cycles. Nonetheless, the distinctive vertical changes and spatial changes in river water isotopes in Tajikistan can provide vital information about water resources.

2. Methods

2.1. Research area

About 93% of Tajikistan's land surface is composed of mountains or plateaus, and more than half of its terrain is over 3000 m. Based on the elevation of the terrain, Tajikistan can be roughly divided into four geographical regions: the northern mountains and basins, mainly belonging to the Tianshan Mountains; the southwest with its lower-altitude valleys; the central mountainous region; and the Pamir plateau in the east. About 6% of Tajikistan is covered by glaciers, with an ice volume reaching 500 km^3 (Makhmadaliev, 2002). These glaciers are mainly in the central and eastern Pamir, with some located in northwestern Tajikistan. The large number of glaciers in the high mountains and plentiful snowfall in winter and spring provide sufficient water resources for the whole country. In the winter, there is heavy precipitation in the form of snow, especially in central Tajikistan. However, the summer is dry and hot in the middle and west. This seasonality affects annual river runoff. As the climate warms from March to May, glaciers and snow begin melting and river runoff increases, reaching a maximum in the period from June to August.

The main moisture supply to the region comes from westerlies, southwestern cyclones and sometimes Caspian Sea and Aral Sea evaporation. The eastern Murghab River and East Pamir Plateau rainfall are concentrated in the period from May to August. Unlike in the east, precipitation in the west of Tajikistan is highest during winter and spring (Fig. 1a), owing to the relief impact of the Siberian/Tibetan high-pressure systems during winter. Summer and autumn are the dry seasons over southwest central Asia. The seasonal precipitation patterns at meteorological stations in western Tajikistan such as at Dushanbe (Fig. 1b), imply a typical Mediterranean climate with concentrated winter and spring precipitation. Eastwards, the seasonal precipitation pattern gradually changes to a summer maximum precipitation pattern as at Carlakul in the Murghab region, an area typically controlled by westerlies (Aizen et al., 2009). From west to east, precipitation amounts decline significantly, especially during winter and spring, and there is an apparent shift from winter and spring precipitation patterns to summer precipitation patterns. The changes in precipitation recorded at six meteorological stations are also reflected in the climate record as listed in Table 1 (precipitation) and Table 2 (air temperature).

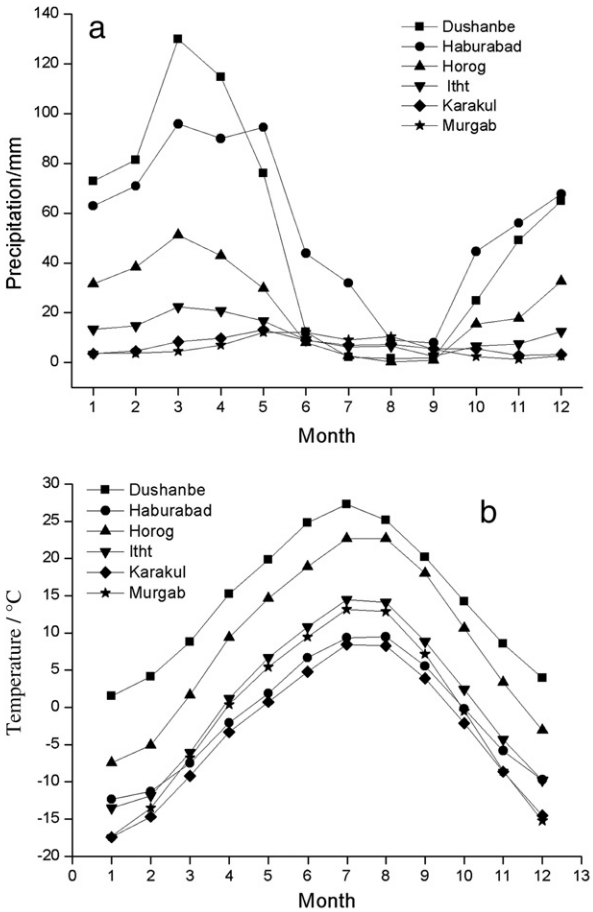

Fig. 1. Average monthly precipitation (a) and monthly air temperature (b) at Tajikistan meteorological stations.

2.2. Water sampling and measurement

A total of 72 river water samples were collected in August 2011, and 12 samples in May 2010, along a west–east transect. One sample was taken from the River Kraft, 14 from the River Vakhsh, 52 from the River Panj, and five from the River Karakul (Fig. 2). These rivers have sources in different parts of Tajikistan and thus receive different water supplies. The River Kraft originates north of Dushanbe, and is the capital's main water resource. The River Vakhsh originates from the glaciers surrounding Peak Somoni and from a northeastern Tianshan glacier. The River Panj rises in western Pamir and the surrounding glaciers. The River Karakul originates from a western Tianshan glacier (Makhmadaliev, 2002). We used a portable GPS to record the location of sampling sites. These samples were sealed in 15 ml bottles during the fieldwork, then frozen in the laboratory until measurement. All water samples were measured with a Picarro-2130i Liquid Water Analyzer at the Key Laboratory of Tibetan Plateau Environment Changes and Land Surface Processes, part of the Institute of Tibetan Plateau Research of the Chinese Academy of Sciences in Beijing. The value of each isotopic ratio is expressed as *per mil* relative to Vienna Standard Mean Ocean Water 2(VSMOW2) and with precision of$\pm$0.15‰ for δ^{18}O, and$\pm$0.4‰ for δD. The measured results are listed in Table 3.

3. Results and discussion

3.1. The relation between δ^{18}O and δD in river water

We first investigated the relation between the δ^{18}O and δD of these Tajikistani river water samples. The results showed that the values of δ^{18}O in Tajikistan river water change from −10.6‰ to −17.8‰, with an average of −14.6‰, and that the δD ranges from −69.1‰ to −132.7‰, with an average of −104.5‰.

We noticed a close relation between δ^{18}O and δD in river water across Tajikistan, thus: $\delta D = 9.26\delta^{18}O + 30.8$ ($R^2 = 0.98$;n = 72). Unlike precipitation isotopes, river water might experience other processes resulting in isotope fractionation, such as post-depositional snow processes and further evaporation, which may modify the relation between δ^{18}O and δD. Earlier work on snow melting processes has shown that evaporation and refreezing result in a slope decrease, and melting has no clear impact on water isotope composition (Zhou et al., 2007). Therefore, the high observed slopes in the δ^{18}O and δD correlation are less likely to be related to snowmelt and refreezing. On average, the evaporation water line is located under the GMWL (Craig, 1961; Rozanski et al., 1993), dependent upon the degree of evaporation (Gonfiantini, 1986). However, the scatter points in Fig. 3 are almost above the GMWL, implying that river water is not significantly affected by land surface evaporation enrichment. However, slight differences exist between samples from different parts of Tajikistan. To compare this spatial variation, we separated these samples into two groups, based upon the topography of the sampling sites. To the west of roughly 71.5°E, consisting of plain and low mountain topography, our sampling elevation was below 2000 m asl. To the east of the dividing line are the high mountain and glacier regions, and our sampling site elevation was therefore usually in the range between 2000 m and 5000 m. This method allowed us to present separately the relation between river water δ^{18}O and δD values in the two regions (Fig. 3). The linear correlation between δ^{18}O and δD is $\delta D = 8.5\delta^{18}O + 21.5$ ($R^2 = 0.99$;n = 28) in the west, and $\delta D = 9.5\delta^{18}O + 30.8$ ($R^2 = 0.98$;n = 44) in the east. We found a slight bias in the relation to the GMWL. In the west, the scatter points are above the GMWL and, in the east, the scatter points are within or below the GMWL. In the west, where there is a lack of summer precipitation, glacier meltwater provides the main river water supply. This yields a less fractionated river water isotope. Local evaporation can change the relation between δ^{18}O and δD in residual water. This can result in a linear relation between δ^{18}O and δD values under the GMWL. However, this is not the case in west Tajikistan, as the distribution of δ^{18}O and δD values there is high above the GMWL (Fig. 3), implying that the river water in west Tajikistan is not obviously affected by local evaporation. The higher *d*-excess in that region also supports this conclusion, as discussed in the next section. In the eastern region, we observed a linear relation in δ^{18}O–δD values in river water slightly below the GMWL. This phenomenon is probably associated with the higher summer precipitation in that region, which may cause higher land surface evaporation. The low *d*-excess in that region also supports this implication. Nevertheless, the observed δ^{18}O–δD relation in Tajikistani river water did show a marked bias in relation to the GMWL. This result shows that the river water did not experience significant isotope fractionation by evaporation.

3.2. Vertical lapse rate of δ^{18}O values in Tajikistani river water

River water isotopes show an apparently consistent change with altitude, due mainly to the precipitation isotope "altitude effect". Fig. 4 shows the longitudinal change in river water δ^{18}O values compared with sampling altitude. The open circles are for 2010 river water samples and the solid circles for 2011 samples (Fig. 4). There are 12 samples from 2010, and their δ^{18}O variations overlap with the 2011 results *vis-à-vis*

Table 1
Average monthly precipitation (in mm) at Tajikistan meteorological stations.

Weather station	Month											
	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec
Dushanbe	72.9	81.4	129.9	114.6	76.0	12.3	2.2	1.6	1.9	24.9	49.1	64.8
Haburabad	63.0	70.9	95.8	90.0	94.5	43.9	31.9	9.0	7.9	44.7	56.0	67.8
Horog	31.6	38.3	51.2	42.8	29.8	8.1	2.7	0.2	1.0	15.5	17.9	32.7
Itht	13.4	14.8	22.5	20.8	16.7	9.2	6.2	6.6	2.6	6.6	7.5	12.5
Karakul	3.5	4.6	8.4	9.7	13.1	8.9	7.0	7.3	5.6	5.6	2.9	3.3
Murgab	3.8	3.7	4.5	7.0	12.1	12.2	9.1	10.4	5.4	2.4	1.4	2.7

Table 2
Average monthly air temperature (°C) at Tajikistan meteorological stations.

Weather station	Month											
	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec
Dushanbe	1.6	4.1	8.8	15.2	19.8	24.8	27.2	25.2	20.2	14.2	8.6	4.0
Haburabad	−12.3	−11.3	−7.5	−2.1	1.9	6.7	9.4	9.5	5.5	−0.2	−5.8	−9.7
Horog	−7.4	−5.0	1.7	9.4	14.7	18.9	22.7	22.7	18.0	10.6	3.4	−3.1
Itht	−13.5	−11.9	−6.1	1.2	6.7	10.8	14.5	14.1	8.8	2.4	−4.3	−9.8
Karakul	−17.4	−14.7	−9.2	−3.3	0.7	4.8	8.5	8.3	3.9	−2.1	−8.6	−14.5
Murgab	−17.3	−13.5	−6.8	0.4	5.4	9.5	13.2	12.9	7.2	−0.5	−8.5	−15.2

longitude. Thus, we believe that, although these samples are from different time periods, they can nonetheless reflect the general spatial trend of river water $\delta^{18}O$ values in Tajikistan. A roughly decreasing trend of river $\delta^{18}O$ with increasing altitude characterizes the west–east spatial variation. The sampling altitude varies in the range from 863 m to 4973 m, with the corresponding river water $\delta^{18}O$ values ranging from −10.6‰ at 69°E to −17.8‰ at 73.25°E.

Fig. 5 shows the relation of observed river water $\delta^{18}O$ values to altitude at which the river waters are sampled. The relation between river water $\delta^{18}O$ and altitude is significant, with an overall correlation between $\delta^{18}O$ and altitude of $\delta^{18}O = -0.0011h - 11.74$ ($R^2 = 0.46$, $n = 72$). In west Tajikistan, river water $\delta^{18}O$ values show greater changes in magnitude within a lower altitudinal range (dot in Fig. 5). This is because the river waters concerned originated from different altitudes in the mountainous regions, thus bearing different isotope signals without any necessary linkage to sampling elevation.

In eastern Tajikistan, we split the river water samples into two groups: main stream and tributary stream. The presence of river water $\delta^{18}O$ in the main streams shows a much more consistent relation thus: $\delta^{18}O = -0.0013h - 12.35$ ($R^2 = 0.88$, $n = 11$), when compared with that in the tributary streams ($\delta^{18}O = -0.0009h - 12.51$ ($R^2 = 0.54$, $n = 43$) (excluding data no. 136)), revealing complex water resources therein. The overall $\delta^{18}O$ vertical lapse rate of 0.11‰/100 m is quite close to that of 0.13‰/100 m from the main stream in east Tajikistan. The observed river water $\delta^{18}O$ vertical lapse rate of 0.09–0.13 is apparently lower than in the global average of 0.28‰/100 m (Poage and Chamberlain, 2001). Although the lower water $\delta^{18}O$ vertical lapse rate in precipitation has been found in the southern Himalaya (0.15‰/100 m: Wen et al., 2012), river water $\delta^{18}O$ lapse rates display an extensive variation ranging from 0.18 to 0.29‰/100 m in the Nepal Himalaya (Garzione et al., 2000), and 0.36‰/100 in the Tibetan Himalaya (Wen et al., 2012). However, the lapse rate 0.09–0.13‰/100 observed in the Tajikistani Pamir is still the lowest ever found around the

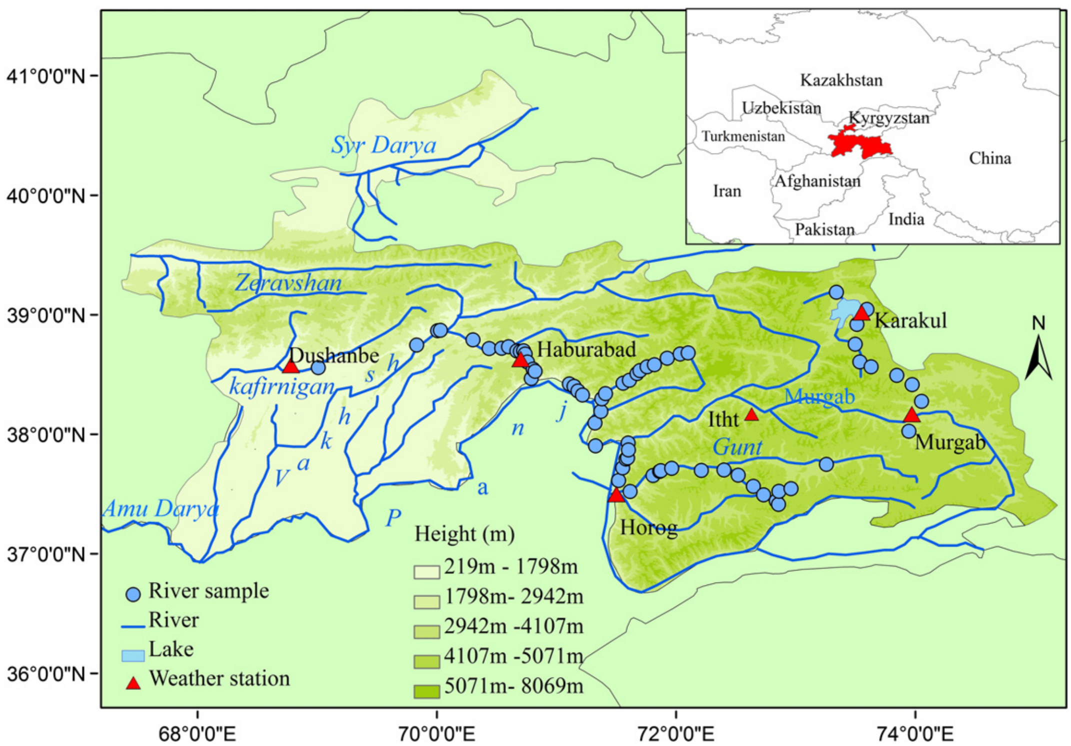

Fig. 2. River water sampling sites and meteorological stations in Tajikistan. Inset is the geographical location of Tajikistan.

Table 3
Geographical locations of river water samples in Tajikistan and measured isotope results.

Sample identification	Catchment	Sampling date	Lat	Lon	Elevation(m)	$\delta^{18}O$ (‰)	δD (‰)	*D*-excess (‰)
S-100	Kraft	2010-5-2	38.6	69.0	863	−11.0	−70.9	16.9
S-101	Vakhsh	2010-5-2	38.7	69.8	1067	−14.2	−99.3	14.5
S-102	Vakhsh	2010-5-2	38.7	69.8	1067	−10.9	−71.9	15.2
S-103	Vakhsh	2010-5-2	38.9	70.0	1167	−14.4	−101.1	13.9
S-104	Vakhsh	2010-5-2	38.9	70.0	1155	−14.6	−102.0	14.8
S-105A	Vakhsh	2010-5-2	38.9	70.0	1155	−14.1	−97.0	15.4
S-105B	Vakhsh	2011-8-5	38.8	70.3	1480	−10.9	−71.2	16.3
S-106	Vakhsh	2010-5-2	38.7	70.4	1553	−13.9	−97.0	14.5
S-107	Vakhsh	2011-8-6	38.7	70.5	1664	−10.6	−69.1	15.4
S-108	Vakhsh	2011-8-6	38.7	70.6	1676	−14.1	−97.2	15.5
S-109	Vakhsh	2011-8-6	38.7	70.7	1905	−12.3	−83.7	14.7
S-110	Vakhsh	2010-5-2	38.7	70.7	2085	−12.1	−83.0	13.9
S-111	Vakhsh	2011-8-6	38.7	70.7	2181	−12.1	−82.2	14.4
S-113	Vakhsh	2011-8-6	38.6	70.7	3111	−12.5	−84.2	15.5
S-112	Vakhsh	2011-8-6	38.7	70.7	2508	−12.6	−86.3	14.8
S-114	Panj	2011-8-6	38.6	70.8	2834	−12.5	−86.0	13.9
S-117	Panj	2011-8-6	38.5	70.8	1427	−12.4	−83.1	15.8
S-115	Panj	2011-8-6	38.5	70.8	1894	−12.2	−81.9	15.6
S-116	Panj	2011-8-6	38.5	70.8	1687	−12.4	−83.2	15.8
S-118	Panj	2011-8-6	38.4	71.1	1487	−13.1	−89.0	15.9
S-119	Panj	2011-8-7	38.4	71.1	1424	−12.9	−88.7	14.2
S-120	Panj	2011-8-7	38.4	71.2	1482	−13.4	−92.0	15.3
S-121	Panj	2011-8-7	38.3	71.2	1507	−15.1	−108.2	12.5
S-134	Panj	2011-8-9	38.1	71.3	1705	−15.2	−109.7	12.1
S-135	Panj	2011-8-9	37.9	71.3	1987	−15.4	−111.4	11.8
S-133	Panj	2011-8-9	38.2	71.4	1660	−15.0	−105.3	14.4
S-122	Panj	2010-5-2	38.3	71.4	1569	−14.7	−102.5	15.3
S-123	Panj	2011-8-7	38.3	71.4	1693	−12.3	−83.7	15.0
S-141	Panj	2011-8-9	37.6	71.5	2056	−14.9	−107.3	11.6
S-140	Panj	2011-8-9	37.7	71.6	2071	−14.0	−99.3	12.4
S-124	Panj	2011-8-7	38.4	71.6	1854	−13.8	−94.3	16.3
S-139	Panj	2011-8-9	37.8	71.6	2029	−14.8	−106.8	11.8
S-138	Panj	2011-8-9	37.8	71.6	2035	−14.4	−102.8	12.8
S-136	Panj	2011-8-9	37.9	71.6	2010	−16.5	−120.9	11.2
S-137	Panj	2011-8-9	37.9	71.6	2019	−14.6	−103.4	13.4
S-125	Panj	2010-5-2	38.5	71.6	1958	−14.0	−98.0	14.3
S-142	Panj	2011-8-1	37.5	71.6	2219	−15.8	−115.6	11.0
S-126	Panj	2010-5-2	38.5	71.7	1963	−14.9	−103.7	15.4
S-127	Panj	2010-5-2	38.5	71.7	2043	−14.8	−102.4	15.8
S-128	Panj	2011-8-7	38.6	71.8	2082	−14.0	−97.4	14.9
S-143	Panj	2011-8-1	37.7	71.8	2551	−14.1	−99.3	13.2
S-129	Panj	2011-8-7	38.6	71.8	2130	−14.8	−104.2	14.4
S-144	Panj	2011-8-1	37.7	71.9	2585	−15.9	−116.5	10.5
S-145	Panj	2011-8-1	37.7	71.9	2616	−14.4	−101.1	13.8
S-130	Panj	2011-8-7	38.6	71.9	2293	−14.9	−105.0	14.1
S-147	Panj	2011-8-1	37.7	72.0	2701	−16.1	−118.3	10.2
S-146	Panj	2011-8-1	37.7	72.0	2701	−15.2	−107.5	13.9
S-131	Panj	2011-8-8	38.7	72.0	2507	−14.4	−100.4	14.6
S-132	Panj	2011-8-8	38.7	72.1	2646	−15.4	−108.3	14.6
S-149	Panj	2011-8-1	37.7	72.2	3006	−16.5	−122.7	9.0
S-148	Panj	2011-8-1	37.7	72.2	3006	−16.1	−117.1	11.8
S-151	Panj	2011-8-1	37.7	72.4	3193	−16.4	−122.9	8.2
S-150	Panj	2011-8-1	37.7	72.4	3193	−16.0	−121.5	6.7
S-152	Panj	2011-8-1	37.7	72.5	3404	−17.0	−126.7	9.5
S-153	Panj	2011-8-1	37.6	72.6	3695	−16.4	−120.6	10.6
S-154	Panj	2011-8-1	37.5	72.7	3928	−15.6	−118.5	5.9
S-156	Panj	2011-8-1	37.4	72.8	4523	−16.9	−125.2	9.9
S-155	Panj	2011-8-1	37.4	72.9	4973	−17.8	−131.2	11.3
S-157	Panj	2011-8-1	37.5	72.9	4144	−15.9	−119.8	7.7
S-158	Panj	2011-8-1	37.5	73.0	4039	−16.5	−125.7	6.1
S-159	Panj	2011-8-1	37.7	73.3	3916	−15.3	−117.6	4.9
S-175	Karakul	2011-8-1	39.2	73.3	3964	−17.7	−132.7	9.2
S-169	Karakul	2011-8-1	38.8	73.5	4089	−16.8	−126.3	7.9
S-170	Karakul	2011-8-1	38.9	73.5	3988	−15.4	−112.8	10.6
S-168	Karakul	2011-8-1	38.6	73.5	4266	−17.5	−129.3	10.5
S-173	Karakul	2011-8-1	39.0	73.6	3864	−14.9	−108.1	11.1
S-167	Panj	2011-8-1	38.6	73.6	4413	−16.4	−125.0	6.3
S-166	Panj	2011-8-1	38.5	73.8	4022	−17.4	−129.1	9.9
S-160	Panj	2011-8-1	38.0	73.9	3813	−14.8	−110.6	7.6

Table 3 *(continued)*

Sample identification	Catchment	Sampling date	Lat	Lon	Elevation(m)	$\delta^{18}O$ (‰)	δD (‰)	*D*-excess (‰)
S-165	Panj	2011-8-1	38.4	74.0	3878	−15.4	−113.0	9.9
S-164	Panj	2011-8-1	38.4	74.0	3878	−14.4	−104.8	10.5
S-163	Panj	2011-8-1	38.3	74.1	3725	−17.0	−127.1	9.1

Third Pole region. It is believed that monsoonal convection induces the lower lapse rate in the southern Himalaya. In the Pamir region, the lower observed $\delta^{18}O$ vertical lapse rate in river water is probably related to specific topography and water resources. The Pamir mountain region is made up of sharp mountains and deep valleys. The deep accumulated snow found in the valleys is a mixed body of snow from different altitudes; these melt between May and August, limiting the isotope selection. Such topography favors a gradual melting of snow and ice across a relatively narrow altitudinal range as the temperature increases progressively from spring to summer. This constrains the variation range of water $\delta^{18}O$, and leads to an observed low river water $\delta^{18}O$ vertical lapse rate.

3.3. Spatial variation of *d*-excess in river water

D-excess, defined by Dansgaard as $d = \delta D - 8\delta^{18}O$ (Dansgaard, 1964), has an average value of about 10 in global precipitation. *D*-excess in precipitation is related to the kinetic fractionation during water evaporation. Humidity relative to saturation at sea surface temperature and wind speed are the major controlling factors on precipitation *d*-excess. Therefore, *d*-excess bears information about the climate in the vapor source region. Further evaporation can modify surface water *d*-excess subsequently. *D*-excess in Tajikistan river water is in the range from 4.9‰ to 16.9‰ (Fig. 6), with an average value of 12.5‰, which is slightly higher than the global average precipitation. There is an apparently spatial decreasing trend in river water *d*-excess from west to east in Tajikistan. Roughly, river water *d*-excess is in the range from 10‰ to 17‰ in the west and between 4‰ and 12‰ in the east (Fig. 6).

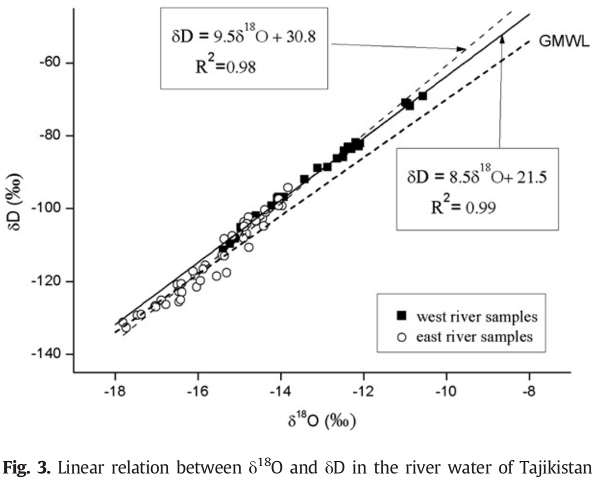

Fig. 3. Linear relation between $\delta^{18}O$ and δD in the river water of Tajikistan as compared with the Global Meteoric Water Line (GMWL). The solid line indicates the linear trend of samples from the west, while the shallow dashed line indicates samples in the east. The dark dashed line is the GMWL of $\delta D = 8\delta^{18}O + 10$.

With regard to the possible factors that can affect the spatial change in river water *d*-excess in Tajikistan, we detected an inverse trend between the spatial river water *d*-excess and elevation. The work shows that *d*-excess increases with altitude as in Antarctica (Qin et al., 1994). Simulation also shows an increasing trend of *d*-excess in vapor with altitude, in comparison with decreasing vapor $\delta^{18}O$ (Bony et al., 2008). Therefore, the spatial decrease in river water *d*-excess in Tajikistan is less likely to be derived from the increase in altitude.

The spatial shift of river water *d*-excess implies a shift of different moisture sources. The higher *d*-excess in river water in west Tajikistan reflects the influence of moisture from Mediterranean, and also from Caspian Sea and Aral Sea. The divide is approximately along the peak of Pamir (Peak I Somoni, with an elevation of 7495 m). Earlier work has shown that the precipitation near to the Mediterranean Sea and other land-bound seas (Aral Sea, Caspian Sea, Black Sea) is characterized by higher *d*-excess (Gat and Carmi, 1970). This moisture can obviously give rise to higher *d*-excess in local precipitation and, thereby, in river water. Fig. 7 shows the obviouslyhigher annual average *d*-excess in and around Kabul. Average *d*-excess in the Fedchenko shallow ice cores in Tajikistan is 17.8‰, which is far higher than the global average *d*-excess in precipitation (Aizen et al., 2009). The lower *d*-excess in river water of east Tajikistan is consistent with a non-Mediterranean source. The different moisture sources are also supported by the apparent seasonal precipitation patterns between the two sides (Fig. 1a).

Another reason for the spatial change of river water *d*-excess is the seasonal precipitation patterns between west and east of Tajikistan. On a broad scale, seasonal *d*-excess in precipitation is associated with evaporation in the vapor sources. Higher relative humidity over oceanic source regions gives rise to lower *d*-excess in precipitation, and lower relative humidity in winter leads to higher *d*-excess in precipitation in the northern hemisphere (Jouzel et al., 1997). Thus, the seasonal precipitation

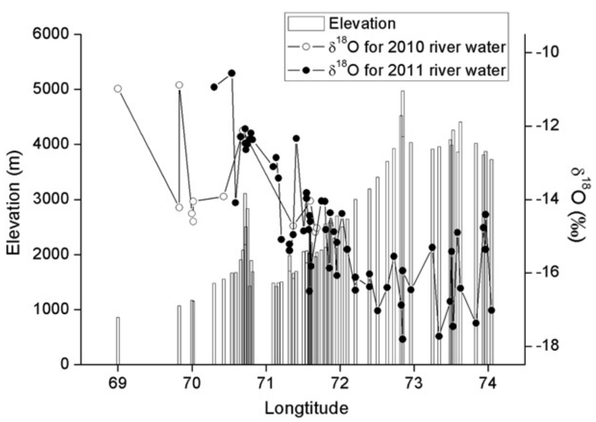

Fig. 4. Variations in $\delta^{18}O$ in river water with sampling elevation in Tajikistan.

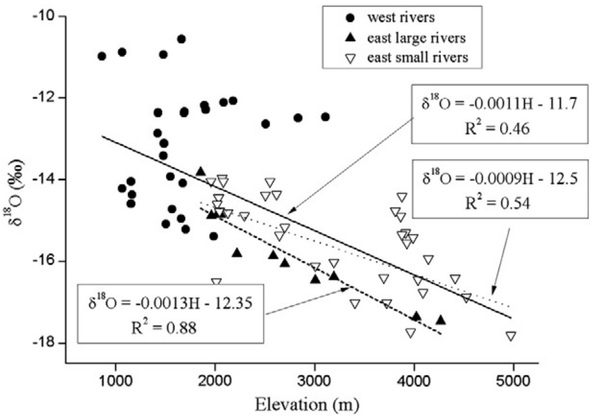

Fig. 5. Relation between river water $\delta^{18}O$ and elevation in Tajikistan. Solid line is the trend line for all samples, dark dashed line for samples from the large rivers in the east, and shallow dashed line for samples from small rivers in the east.

pattern has a significant impact on precipitation *d*-excess. In the Himalayan case, there is significantly higher *d*-excess in river water when compared with the large region of the southern Tibetan Plateau. *D*-excess values of over 20‰ are related to the higher precipitation amounts in winter and spring with higher precipitation *d*-excess in those two seasons, while there is almost no winter and spring precipitation in most parts of the southern Tibetan Plateau, leading to lower monsoon precipitation *d*-excess in precipitation and river water (Tian et al., 2001, 2007).

There is a significant precipitation pattern shift from west to east in Tajikistan. At stations in the middle and west of Tajikistan (Fig. 1a), precipitation falls mainly in winter and spring, and there is a definite lack of summer rainfall. However, this precipitation pattern changes to a summer-dominated seasonal precipitation pattern in east Tajikistan. Because there is no directly observed precipitation isotope data in Tajikistan, we present seasonal precipitation *d*-excess patterns at stations surrounding Tajikistan based on GNIP work (Fig. 7). Given the different precipitation patterns, the two stations at Kabul and Urumqi show a higher winter and a lower summer *d*-excess in precipitation, in agreement with the seasonal patterns in the northern hemisphere. If we consider both kinds of seasonality in precipitation and *d*-excess, we find a higher annual

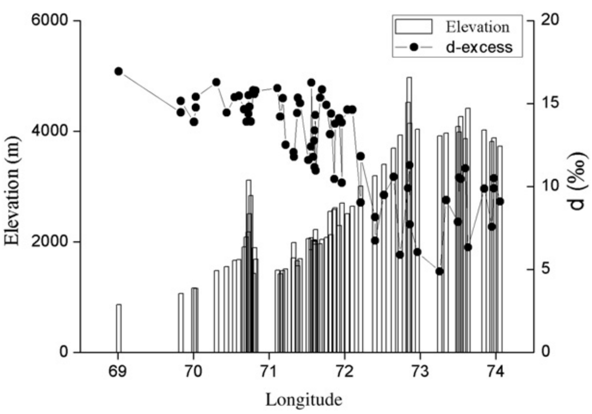

Fig. 6. Longitudinal variation of river *d*-excess in Tajikistan compared to changes in sampling elevation.

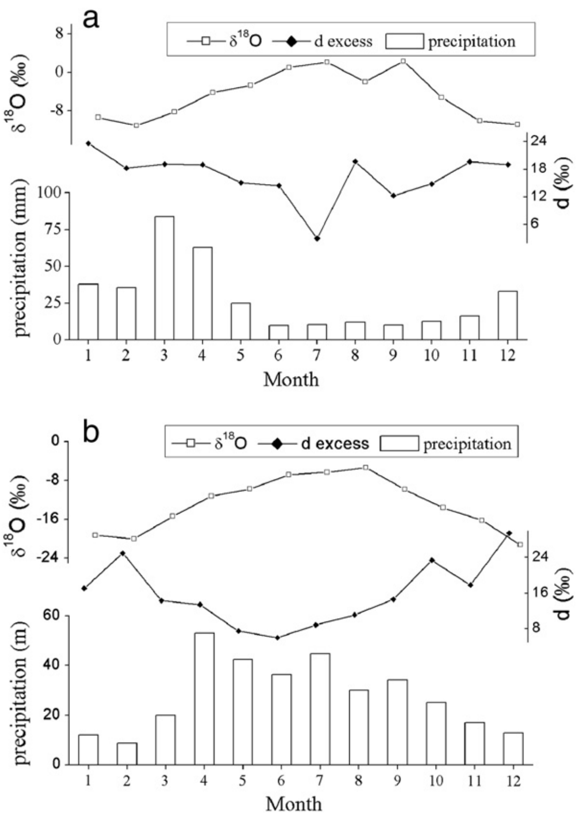

Fig. 7. Seasonal change in $\delta^{18}O$, *d*-excess and monthly precipitation amount at Kabul (a) and Urumqi (b).

precipitation *d*-excess in the west of Tajikistan due to the apparent winter and spring precipitation, and a lower *d*-excess in east Tajikistan.

However, direct observation of precipitation isotopes in different parts of Tajikistan can greatly enhance our understanding of the different moisture origins and their effect on surface water isotopes in Tajikistan.

4. Conclusions

Tajikistan has the highest peaks, longest glaciers and the greatest glacial meltwater resources in Middle Asia. The potential influence of climate change on water sources, the ice core record from that region and the paleo-plateau uplift history of Pamir necessitate an understanding of the water isotopes in moisture sources and the hydrological cycle. However, research into the isotope hydrology in Tajikistan is very limited for a variety of reasons. We carried out river water isotope work in Tajikistan in an effort to understand the spatial change in river water isotopes as well as their altitudinal change.

Water $\delta^{18}O$–δD relations can provide some hints on surface water evaporation. Here we present a close relation ($\delta D = 9.26\delta^{18}O + 30.8$ ($R^2 = 0.98$;n = 72)) in Tajikistan river water. Although there is a slight difference between east and west regions, the scatter dots of δD versus $\delta^{18}O$ in river water are generally above the Global Meteoric Water Line (GMWL), implying that the river water has not experienced significant isotope fractionation by evaporation processes. These results accord with the river water sources in Tajikistan, where snow and ice melting is an important source of river water.

We found an obvious decreasing change in river water $\delta^{18}O$ by altitudinal sampling. The river water $\delta^{18}O$ decreases from −10.6‰ to −17.8‰, while the sampling altitude increases from 863 m to 4973 m. The observed river water $\delta^{18}O$ vertical lapse rate is 0.09–0.13‰/100 m, apparently lower than the global average of 0.28‰/100 m. The lower water isotope lapse rate is probably related to specific topography in the mountain regions of Tajikistan and the seasonal precipitation pattern, together with a gradual melting rate of snow and ice from lower to higher altitudes.

D-excess is usually used as an indicator of moisture source because of its sensitivity to the condition of water evaporation, but it may also be used to identity different hydrological cycle processes. Our work shows that the river water *d*-excess indicates a significant longitudinal variation in Tajikistan. Roughly, river water *d*-excess varies within the range 10‰–17‰ in the west, but decreases to between 4‰ and 12‰ in the east. This spatial change is not related to changes in altitude but, rather, to the possibility of different moisture sources contributing and seasonal precipitation isotope shift from the west to the east of Tajikistan. Higher *d*-excess implies the influence of moisture from the Mediterranean or other in-land seas, in association with higher *d*-excess. However, lower *d*-excess in east river water is consistent with a non-Mediterranean source. Another reason is the shift of seasonal precipitation patterns. From this west to east pathway, seasonal precipitation shifts from a spring-maximum pattern to a summer-maximum one, while precipitation *d*-excess seasonality is consistent in most of the northern hemisphere and the east may be more influenced by non-Mediterranean moisture. This yields a spatial change as observed in the *d*-excess of the river water in Tajikistan.

However, we also note that large fluctuations were found in the isotopes in our samples, implying that further long-term observation would significantly consolidate any initial conclusions. We also acknowledge that observation of precipitation isotopes is critical for understanding the local water cycle, especially with regard to seasonality and its link to climate change.

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