

Stable isotope systematics of surface water bodies in the Himalayan and Trans-Himalayan (Kashmir) region

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Stable hydrogen (δD) and oxygen ($\delta^{18}O$) isotope ratios of the headwaters of the Indus and its tributaries, surface ice in glaciers, saline and fresh water lakes and thermal springs in the Himalayan and Trans-Himalayan (Kashmir) region are reported. The δD - $\delta^{18}O$ relationship for the river samples shows a slope of 9.12 ± 0.29 which agrees well with the estimate of 8.99 ± 0.33 based on a simple Rayleigh fractionation model. The unique signature of a higher deuterium excess (d) of the 'Western Disturbance' is preserved in these samples. An altitude effect of -0.9 per mil/km is observed in the $\delta^{18}O$ of Indus waters. At a lower altitude (Beas) the altitude effect is almost double, indicating that the altitude effect decreases with elevation in this region.

1. Introduction

Stable isotope studies on the water bodies of the Himalayan region can be useful to model the hydrological cycle and delineate sources of water to various reservoirs in this region. Such studies have been limited (Giggenbach *et al* 1983; Ramesh and Sarin 1992; Bartarya *et al* 1995). To fill this gap we have undertaken a stable isotope study of the headwaters of the Indus and its tributaries, surface ice in glaciers, fresh and saline lakes and thermal springs in the Himalaya and Trans-Himalaya (Kashmir), a hydrometeorologically significant but difficult and inaccessible region. In addition stable isotope measurements of surface water bodies help to interpret stable isotope data on other geological reservoirs (carbonates, organic matter, etc.) in terms of palaeoclimatic changes.

The area sampled (figure 1) comprises two distinct physiographic regions. The Trans Himalayan (Ladakh) region is bounded in the N-E by Tibet and is one of the most elevated regions in the world with no place being less than 3000 m in elevation. The joint

effect of elevation and insolation on snowy mountains produces a unique and extreme climate (Raina 1977). Rainfall is very low (only 116 mm/yr) and the minimum temperature is $-8.5^{\circ}C$ (January) and the maximum $17.4^{\circ}C$ (July) (World Weather Records 1967). From November to March precipitation is in the form of snow (38% of the annual). There is practically no vegetation in this arid region. The Indus source water, glaciers, lakes and spring water samples belong mainly to this region. The other samples are from the lower, Himalayan region. This region is highly vegetated, experiences summer monsoon rain (in contrast to the Trans-Himalaya) and has a range of elevation from 600 m to 3000 m.

2. Sampling

Water samples were collected during August 1991 in clean 30 ml tightly capped polyethylene bottles. Fresh surface ice from the snouts of glaciers was collected, melted and stored in sample bottles. The lids were secured tightly to prevent evaporation

Keywords. Himalaya; springs; stable isotopes; altitude effect.

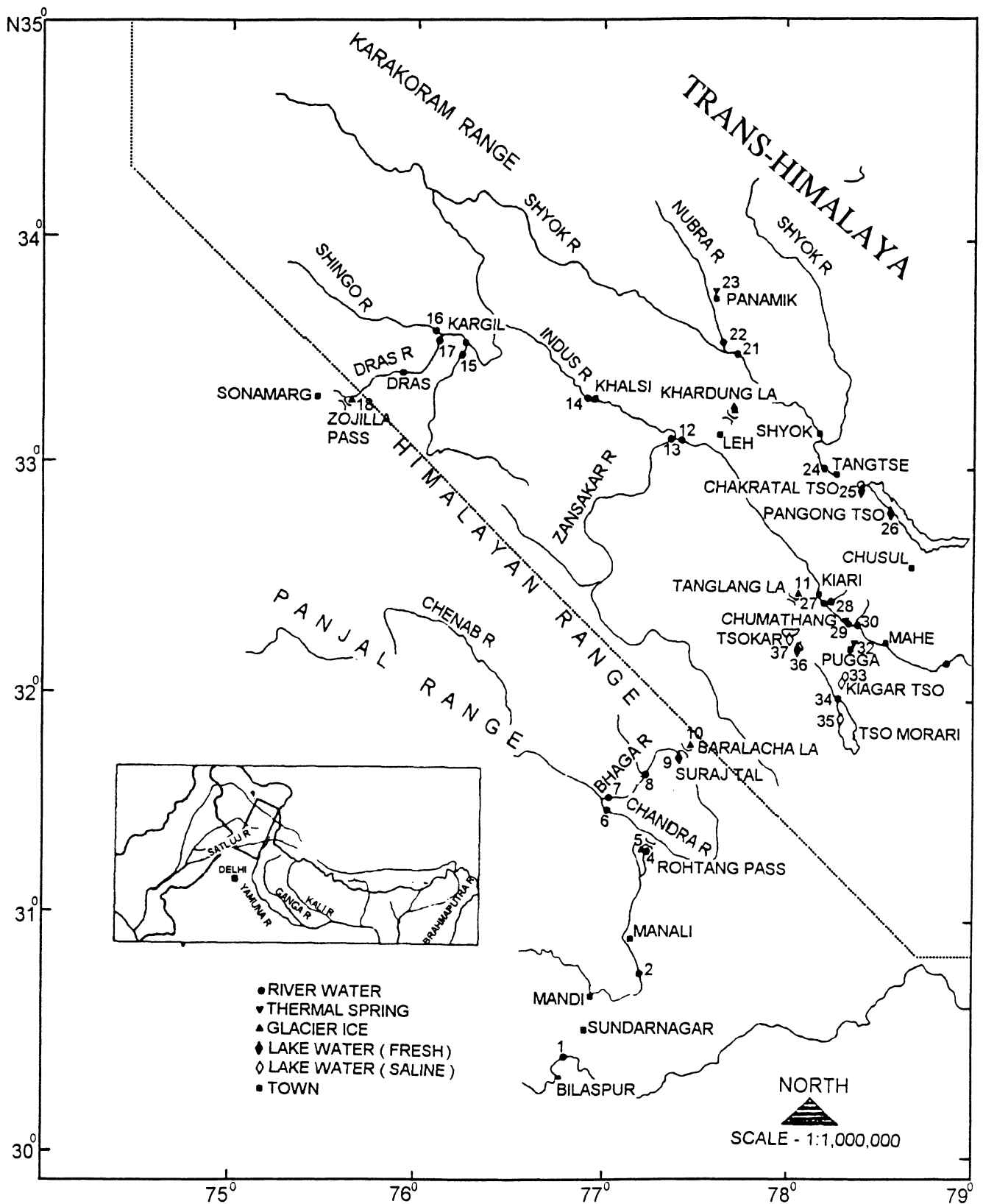


Figure 1. Map of the area sampled; numbers denote sample locations.

and exchange with the atmospheric water vapour. All samples were analysed within a month of collection. Sample numbers and locations are shown in figure 1.

3. Analytical methods

About 2 ml of each sample was equilibrated with tank CO₂ at 25°C for 48 h and the CO₂ then extracted,

purified and let into a VG Micromass 602D mass spectrometer (VG Analytical, Wythenshawe Manchester, UK). $\delta^{18}\text{O}$ was determined with an overall precision of $\pm 0.2\text{‰}$ (Epstein and Mayeda 1953). About $10\ \mu\text{l}$ of each sample was passed through uranium filings kept at 800°C and reduced to hydrogen gas, which was then transferred to 602D VG Micromass mass spectrometer for δD measurement; δD was measured with an overall precision of $\pm 1\text{‰}$ (Friedman 1953). All values are reported relative to V-SMOW (Vienna Standard Mean Ocean Water; Gonfiantini 1981).

4. Results and discussion

The results are presented in table 1. The Indus river and its tributaries have $\delta^{18}\text{O}$ values in the range -11 to -17‰ and δD from -70 to -124‰ , typical of high altitude precipitation. A plot of $\delta\text{D}-\delta^{18}\text{O}$ is shown in figure 2. The 19 river water samples define a best fit line given by

$$\delta\text{D} = (9.12 \pm 0.29)\delta^{18}\text{O} + (31.1 \pm 4.2)$$

with a correlation coefficient (r) 0.98, significant at 0.01 level (P) (dashed line). This regression was

Table 1. δD , $\delta^{18}\text{O}$ and deuterium excess (d) values in (‰) relative to SMOW in rivers, lakes, glaciers and hot springs in the Kashmir Himalaya. Analytical uncertainty $\pm 0.2\text{‰}$ for $\delta^{18}\text{O}$ and $\pm 1\text{‰}$ for δD .

Sample No. (see map)	Description	Altitude (m)	$\delta^{18}\text{O}$	δD	d
Rivers					
1	Sutlej	600	-13.7	-95	14.6
2	Beas	985	-11.1	-71	17.8
6	Chandra	2950	-14.1	-95	17.8
7	Bhaga	2935	-14.7	-102	15.6
8	Darcha	3400	-15.3	-107	15.4
12	Indus	3230	-14.6	-109	7.8
13	Zanskar	3230	-16.1	-114	14.8
14	Indus	3030	-15.2	-108	13.6
15	Suru	2750	-14.8	-104	14.4
16	Shingo	2770	-13.4	-91	16.2
17	Dras	2770	-12.8	-84	18.4
21	Shyok	3260	-16.2	-121	8.6
22	Nubra	3230	-15.5	-113	12.0
24	Tangtse	4945	-15.2	-104	17.6
27	Indus	3560	-15.3	-108	14.4
28	Indus	3630	-14.3	-96	18.4
30	Indus	4060	-15.9	-111	16.2
31	Indus	4130	-15.7	-115	10.6
34	Samdoh	4640	-17.3	-124	14.4
Lakes					
4	Beas Kund (fresh)	3740	-10.3	-65	17.4
9	Suraj Tal (fresh)	4650	-15.7	-110	15.6
25	Chakratral Tso (fresh)	4250	-15.1	-111	9.8
26	Pengong Tso (salt)	4250	-1.4	-42	-30.8
33	Kiagar Tso (salt)	4610	-2.3	-54	-35.6
35	Tso Morari (salt)	4480	-6.3	-62	-11.6
36	Sputtsong Tso (fresh)	3510	-7.9	-75	-11.8
37	Tso Kar (salt)	3510	-6.6	-79	-26.2
Glaciers					
5	Rohtang	3748	-18.1	-128	16.8
10	Balacha La	4650	-15.9	-119	8.2
11	Tanglang La	5210	-24.7	-180	17.6
18	Zozilla	3540	-12.2	-90	7.6
19	Khardung La	5629	-17.2	-127	10.6
20	Khardung La	5649	-15.3	-107	15.4
Hot springs					
23	Ponamik	3320	-17.1	-126	10.8
29	Chumathan	4060	-15.9	-128	-0.8
32	Pugga	4640	-15.0	-126	-6.0

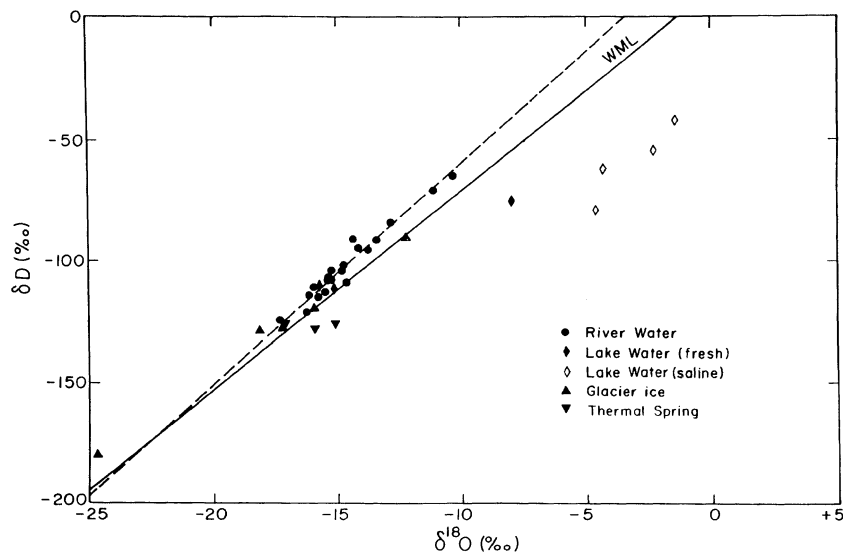


Figure 2. $\delta D - \delta^{18}O$ plot of samples shown in figure 1. Solid line denotes the World Meteoric Water Line (WML). Dashed line is the best fit line for the river water data only.

carried out by weighting the data points with their analytical uncertainties, following the procedure of Williamson (1968). This can be compared with the world meteoric water line (Yurtsever and Gat 1981) (WML, solid line):

$$\delta D = (8.17 \pm 0.28)\delta^{18}O \pm (10.55 \pm 0.64)$$

and that for the Ganges (Ganga) headwaters (Ramesh and Sarin 1992).

$$\delta D = (7.45 \pm 0.23)\delta^{18}O + (8 \pm 2).$$

In the case of Trans-Himalayan rivers sampled, it is clearly seen that both the slope and the intercept are significantly higher than those of WML and Ganges (Ganga). The main reason for the slope being greater

than 8 is that part of the annual precipitation falls as snow in this region and the mean annual temperature is low ($5.5^{\circ}C$). Table 2 shows the monthly precipitation and temperature data for the Trans-Himalayan Ladakh region. It is seen that for five months (November to March) the air temperature is less than or close to $0^{\circ}C$. Precipitation during these months is in the form of snow (about 38% of the total annual precipitation). If we use a simple Rayleigh Fractionation model (Gat 1981) to calculate the slope of the $\delta D - \delta^{18}O$ relationship, in which the rate of change of δ value with respect to logarithm of the fraction of vapour left in the cloud is equal to $(\alpha - 1) \times 10^3\%$, where α is the fractionation factor between liquid water and vapour [this applies for both the isotopes D

Table 2. Model calculation of slope of the $\delta D - \delta^{18}O$ relationship based on meteorological (World Weather Records 1967) and equilibrium fractionation factors (Majoube 1971) between liquid-vapour except for winter months (marked with *) where ice-vapour fractionation is considered.

Month	Precipitation (mm)	Temperature ($^{\circ}C$)	Fractionation factors		Slope
			Deuterium	Oxygen-18	
January*	12	-8.5	1.148	1.0164	9.02
February*	9	-5.5	1.142	1.0160	8.88
March*	12	0.1	1.132	1.0152	8.68
April	7	5.6	1.113	1.0113	10.00
May	7	10.0	1.098	1.0107	9.16
June	4	13.9	1.090	1.0101	8.91
July	16	17.4	1.085	1.0098	8.67
August	19	16.9	1.083	1.0100	8.30
September	12	13.1	1.094	1.0103	9.13
October	7	6.7	1.111	1.0111	10.00
November*	3	0.6	1.129	1.0130	9.92
December*	8	-4.8	1.140	1.0150	9.33
Mean (or total)	116	5.5	-	-	8.99 ± 0.33

and ^{18}O], the slope of the $\delta\text{D}-\delta^{18}\text{O}$ relationship for precipitation is given by $(\alpha_{\text{D}}-1)/(\alpha^{18}\text{O}-1)$. The fractionation factors for the corresponding monthly temperatures are also shown in table 2. The slope is calculated for each month and an average of all the slopes, weighted with precipitation, is found to be 8.99 ± 0.33 , in good agreement with the observed value of 9.12 ± 0.29 . For the period November to March we have used ice-vapour equilibrium fractionation factors while for other months we have used those of liquid water-vapour (Majoube 1971). Thus to a first order approximation, the higher than 8 slope appears to be mainly caused by the low air temperatures and snow precipitation. Rozanski *et al* (1993) note that for stations with substantial contribution of snow precipitation, higher deuterium excess values (discussed below) during winter may result from additional kinetic fractionation during snow formation (Jouzel and Merlivat 1984). This could also lead to a slope greater than 8.

However, stable isotope studies of cold springs and rivers in Tapoban and Badrinath in UP Himalaya (elevation ~ 3000 m) showed a relationship

$$\delta\text{D} = 7.7\delta^{18}\text{O} + 13.0 \quad (r^2 = 0.94, n = 7)$$

(Navada *et al* 1985). In similar high altitude stations elsewhere in the world (e.g. Ecuador 3058 m, Columbia 2547 m and Adis Ababa 2360 m) the slopes are either 8 or slightly less. Therefore our results could be specific to the Ladakh region only.

We examine the deuterium excesses parameter (d) defined as $\delta\text{D}-8\delta^{18}\text{O}$, shown in table 1. As about 38% of the annual precipitation is derived from westerlies ('Western Disturbance') whose source is the Mediterranean Sea (Raina 1977) with a characteristically higher deuterium excess (Gat and Carmi 1970) the river water samples also exhibit higher deuterium excess. The average d value is 14.7 ± 3.1 ‰. All along the path of the Western Disturbance (Iran, Afghanistan, Pakistan and Kashmir) the precipitation is characterised by high deuterium excess (Yurtsever and Gat 1981). In a compilation of deuterium excess data from precipitation samples collected world wide, Rozanski *et al* (1993) found that between 30° and 40°N , most samples show high deuterium excess (in the range 15 to 25 per mil). Thus the high values of deuterium excess in our samples confirm the findings of Rozanski *et al* (1993), who also showed that the increase of the deuterium excess was correlated with the seasonal bias of precipitation towards winter. High deuterium excess values ($\sim 13.00 \pm 3.29$ ‰) have also been reported from Parbati Valley ground and cold spring waters (Giggenbach *et al* 1993).

The enhanced d values could also be due partly to a contribution of moisture from evaporation from local water bodies to the cloud formation. Such local evaporation can occur from fresh and salt water lakes in this arid region. Figure 2 shows that samples from

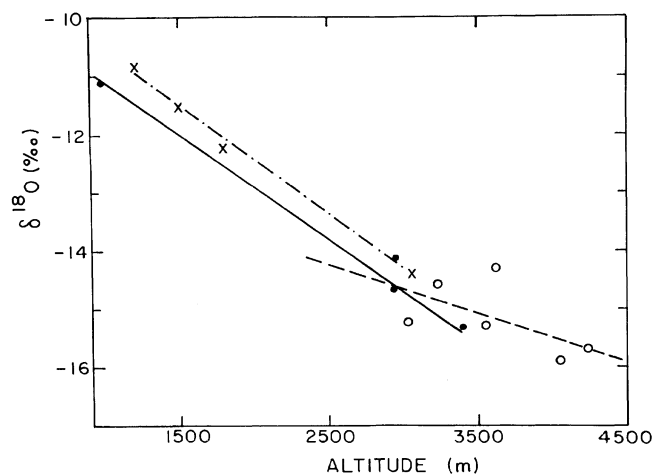


Figure 3. $\delta^{18}\text{O}$ as a function of altitude; solid line – Beas (filled circles); dashed line – Indus (open circles); dashed dotted line – Ganges main channel (crosses) (data from Ramesh and Sarin 1992).

the four saline lakes and one fresh water lake (Sputtsong Tso) plot well below the local meteoric line and WML indicating that these have undergone significant evaporation. The point of intersection of the best fit line for these points with the best fit line for the river water samples gives an approximate estimate of their original isotopic composition. It can therefore be estimated that annual evaporation could be as high as 70% of the annual precipitation in these lakes (estimate based on Rayleigh isotopic fractionation model taking into account the $\delta^{18}\text{O}$ atmospheric water vapour). Further, the six glacier snout ice samples also show a reduced slope (line not shown in figure 2): $\delta\text{D} = (7.29 \pm 0.19)\delta^{18}\text{O} + (0.4 \pm 3.4)$ with $r = 0.99$, $P = 0.01$ suggestive of sublimation of ice and local contribution of moisture to the clouds.

Figure 3 shows a plot of $\delta^{18}\text{O}$ of some of the samples as a function of altitude. As different rivers in this region (Beas, Sutlej and Indus) receive their water from different source regions, it is possible to detect the altitude effect only along the main channel of a single river. For the Indus river (open circles) in the Trans-Himalaya, the equation is

$$\delta^{18}\text{O} = -(0.086 \pm 0.011)\text{alt} - (12.07 \pm 0.41)$$

with $r = -0.58$ $P = 0.01$

where alt is altitude in units of 100 m.

For Beas,

$$\delta^{18}\text{O} = -(0.17 \pm 0.01)\text{alt} - (9.39 \pm 0.16)$$

with $r = -0.99$ $P = 0.01$.

Thus the altitude effect for Indus (Trans Himalaya) is about -0.09 ‰ for 100 m whereas for Beas (Himalaya) it is -0.17 ‰ per 100 m, comparable to that for Ganges (Ramesh and Sarin 1992). Giggenbach *et al* (1983) in an isotope study of hot and cold springs in the Parbati Valley, Himachal Pradesh, report an

altitude effect of -1% per km for cold springs. This is similar in magnitude to the value observed by us for the Indus river data.

It has been shown that the ‘‘altitude effect’’ in Ganges river water samples is about half that in precipitation in the source region of Ganges (Ramesh and Sarin 1992) due to the averaging of the $\delta^{18}\text{O}$ values as the river picks up water at different altitudes during its descent. A similar conclusion is valid for the Beas samples. However, for the Indus samples (higher altitude compared to Beas) the altitude effect is about half that seen for Beas. The reason could be that the altitude effect rapidly decreases with elevation in this region. It has been shown that $\delta^{18}\text{O}$ in fresh snow from Garhwal does not show any altitude effect in the range 4400 to 6050 m (Niewodniczanski *et al* 1981). Our results are consistent with their finding. Another possible explanation would be as follows : Most of the air masses come south and as they are rising on the southern slope of Himalaya, the vapour condenses and produces precipitation. When the top elevation is reached there is not much vapour left (and therefore the northern slope is arid) and the influence of altitude on the condensation process and isotopic composition of precipitation becomes negligible. This phenomenon has also been observed in Gran Canaria, Canary Islands (Gonfiantini *et al* 1976). Therefore the small altitude effect could be mainly due to evaporation during rainfall. According to another view (Navada, personal communication) the Indus during its descent may not be picking up much water and this could be a cause of the reduced altitude effect.

Mayewski and Lyons (1983) reported δD data for 8 snow samples from Zaskar between 5100 and 5500 m elevations. Our glacier samples show similar values (table 1). Giggenbach *et al* (1983) reported δD , $\delta^{18}\text{O}$ values for 3 snow samples from an altitude range 3400 ~ 3800 m in Himachal Pradesh. Their values are enriched by an average of 10% in $\delta^{18}\text{O}$ and 80% in δD compared to the glacier samples reported in table 1 (from the altitude range 3500 to 5600 m) indicating substantial amount of high altitude snow precipitation in our samples. Likewise, the isotopic composition of thermal springs are also lower (by the same magnitude) than those of the hot springs from Parbati Valley (Giggenbach *et al* 1983) and Nepal (Grabczak and Kotarba 1985). This indicates that the former are being recharged by precipitation from higher altitudes.

5. Conclusion

We have reported δD , $\delta^{18}\text{O}$ data for the Himalayan and Trans-Himalayan water bodies including Indus river and its tributaries. The Indus waters have a higher than normal slope in the δD - $\delta^{18}\text{O}$ plot as the precipitation in this region is partly in the form of

snow and the mean annual temperature is in general low. The theoretically calculated slope is in good agreement with the observed value. Also, these samples have in general higher deuterium excess values indicating that they preserve the signature of ‘Western Disturbance’. The altitude effect in this region decreases with elevation.

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