Vol. 53: 1–12, 2012
doi: 10.3354/cr01090

CLIMATE RESEARCH
Clim Res

Published online May 3

Atmospheric circulation change in the central Himalayas indicated by a high-resolution ice core deuterium excess record

Hongxi Pang[1], Shugui Hou[1,2,*], Susan Kaspari[3], Paul Mayewski[4], Douglas Introne[4], Valérie Masson-Delmotte[5], Jean Jouzel[5], Zhenchao Li[2], Yuanqing He[2], Sungmin Hong[6], Dahe Qin[2]

[1]Key Laboratory of Coast and Island development of Ministry of Education, School of Geographic and Oceanographic Sciences, Nanjing University, Nanjing 210093, China

[2]State Key Laboratory of Cryospheric Science, Cold and Arid Regions Environmental and Engineering Research Institute, Chinese Academy of Sciences, Lanzhou 730000, China

[3]Department of Geological Sciences, Central Washington University, Ellensburg, Washington 98926, USA

[4]Climate Change Institute and Department of Earth Sciences, University of Maine, Orono, Maine 04469, USA

[5]IPSL/Laboratoire des Sciences du Climat et de l'Environnement (LSCE), CEA/CNRS/UVSO, CEA Saclay, 91191 Gif-sur-Yvette, France

[6]Department of Ocean Sciences, College of Natural Science, Inha University, 253, Yonghyun-dong, Nam-gu, Incheon 402-751, Korea

ABSTRACT: Continuous measurements of both δD and $\delta^{18}O$ were performed along a 108.8 m ice core recovered from the East Rongbuk Glacier on the northeast saddle of Mt. Qomolangma (Everest) (28.03° N, 86.96° E, 6518 m above sea level) in September 2002. They provide the first high-resolution historical record of deuterium excess (*d*) in the central Himalayas. In this paper, we focus on *d* variability from 1951 to 2001 and its relationship with large scale atmospheric circulation. The *d* record exhibits significant seasonal variations, with low values in summer and high values in winter, reflecting the atmospheric circulation shift between winter westerlies and the Indian summer monsoon (ISM). The interannual *d* variation is primarily controlled by the ISM moisture transport. An abnormally high *d* value during the period 1960–1964 is linked with the strengthening of winter westerlies, while an anomalously low *d* value during the period 1965–1968 is primarily a result of the migration of the ISM moisture source region, and secondly of surface sublimation. The results show that the ice-core *d* record retrieved from the high Himalayas is a good proxy for changes in atmospheric circulation.

KEY WORDS: Himalayas · Ice core · Deuterium excess · Atmospheric circulation

1. INTRODUCTION

As the 'third pole' of the world, the Tibetan Plateau is a key region for climate change research. The Himalayan region in the southern Tibetan Plateau is dominated by both winter westerlies and the Indian summer monson (ISM). However, the local atmospheric circulation and hydrological processes in the high Himalayas related to the 2 circulation systems are not well understood due to the lack of modern meteorological and climatic observations as well as to its complex topographic conditions. Water stable isotopes, such as deuterium (δD), oxygen-18 ($\delta^{18}O$) and their linear combination, the deuterium excess (*d*), have the potential to provide an effective tracer to better understand atmospheric circulation and mois-

*Corresponding author: Email: shugui@nju.edu.cn

ture origin. Deuterium excess ($d = \delta D - 8\delta^{18}O$) defined by Dansgaard (1964) is indicative of kinetic fractionation during water phase changes (δD and $\delta^{18}O$ are the water isotopic compositions expressed in δ units per mille versus the Vienna Standard Mean Ocean Water [VSMOW]). The slope of 8 is derived from the Global Meteoric Water Line ($\delta D = 8\delta^{18}O + 10$), which describes present day global average of isotopes in precipitation (Craig 1961). The d value is proportional to the effective coefficient of fractionation of water evaporation; it is only slightly changed in the course of equilibrium condensation processes, thus carrying quantitative information about the climate conditions at the oceanic moisture source region (Dansgaard 1964, Craig & Gordon 1965). Merlivat & Jouzel (1979) reported that d is primarily dependent on the mean relative humidity of the air masses formed above the ocean surface, with secondary effects from temperature and wind speed (Uemura et al. 2008). As a result, reconstructing d time series from ice cores offers the potential to estimate how conditions in moisture source regions have varied in the past.

The use of d in reconstructing atmospheric circulation has been widely applied for Antarctic and Greenland ice cores (Vimeux et al. 2001, Masson-Delmotte et al. 2005, Jouzel et al. 2007). However, only a few studies have so far addressed d from ice cores in non-polar regions (Schotterer et al. 1997, Kreutz et al. 2003, Ramirez et al. 2003, Aizen et al. 2005, Vimeux et al. 2008). Notably, a decadal resolution d profile obtained from the Dasuopu ice core from the Himalayas (Thompson et al. 2000) is the single long term (AD 1000 to 2000) record for this region. Herein we present a high-resolution d record over seasonal and inter-annual timescales from an ice core recovered from the East Rongbuk (ER) Glacier on the northeast saddle of Mt. Qomolangma (Everest) and discuss its variability associated with change in atmospheric circulation.

2. SEASONAL ATMOSPHERE CIRCULATION PATTERNS IN THE STUDY AREA

The mountains of the Himalayas and Tibetan Plateau intercept considerable moisture amounts due to the topographic blocking effect. During the wet season (June to September), the ISM transports large amounts of water vapor from the Indian Ocean to the Himalayas and the southern Tibetan Plateau, and brings abundant precipitation into these regions. The 2 dominant monsoon moisture trajectories are from the Indian Ocean across the Arabian Sea to the Tibetan Plateau, and from the Bay of Bengal northward to the central Tibetan Plateau along the Yalongzangbo River valley (Lin & Wu 1990) (Fig. 1a). During the dry

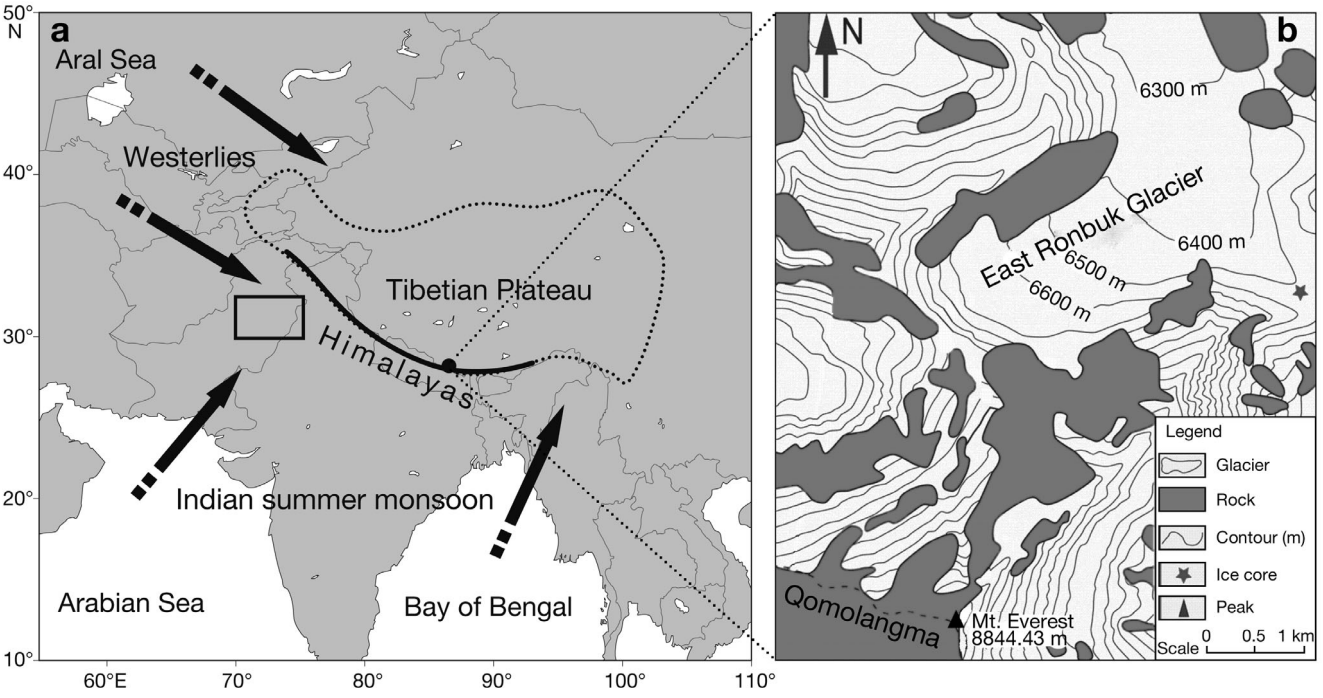

Fig. 1. (a) Dominant atmospheric circulation systems over the study region. (●) Core drill site; rectangle: 'notch' (30–32.5° N, 70–75° E) between western Himalayas and Hindu Kush Mountains; bold line: range of Himalayas; dashed line: general geographical extent of Tibetan Plateau over 3000 m in altitude. (b) ER glacier and ice core drilling site

season (October to May), winter westerlies split into a northern and southern branch around the Tibetan Plateau (Fig. 1a), the latter affecting the Himalayas and southern Tibetan Plateau (Wei & Gasse 1999). In the western Himalayan region, the primary weather system called the 'western disturbances' is a low-pressure system that develops when an upper-level disturbance passes over the notch (30°–32.5° N, 70°–75° E) formed by the Himalayas and Hindu Kush mountains (Fig. 1a), which bring considerable snowfall over the western Himalayan region (Rao 2003, Lang & Barros 2004). Wintertime precipitation in the eastern Himalayas is less than that in the western Himalayas because the influence of the 'western disturbance' on the eastern region is weak.

3. MATERIALS AND METHODS

The ice core, diameter 9.4 cm, 108.8 m long, was drilled in September 2002 using an electromechanical drill in a dry hole. The core was maintained at <–5°C from the time of drilling until analysis. The ER Glacier covers an area of 48.5 km^2 with a length of 14 km (Fig. 1b). Its equilibrium line at 6250 m above sea level is among the highest on Earth (Kang et al. 2005). Repeated GPS surveys in 1998 and 2002 at the ER saddle did not detect horizontal movement of the glacier, indicating minimal flow deformation, yielding an undeformed record. This is supported by the presence of horizontal ice layers in the core. Ice core borehole temperatures ranged from –8.9°C at 10 m to a minimum of –9.6°C at 20 m, then warmed slightly to –8.9°C at the bottom. The combination of high accumulation rates (water equivalent, 50 cm yr^{-1}) and low ice temperature results in the preservation of the isotopic seasonal signal in the ER ice core.

The ice core was shipped frozen to Lanzhou, then to the University of Maine for processing. The core was melted into 3123 discrete samples at 3 to 4 cm resolution using an aluminum melter head with the University of Maine's continuous melter system (Osterberg et al. 2006). The melter head splits the meltwater into 2 different channels; meltwater from the outer portion of the core is collected in an outer channel for isotope analyses while the inner water is used for glaciochemical analysis (Kaspari et al. 2007). Measurements of δD were performed by a Micromass Isoprime mass spectrometer coupled to a Eurovector elemental analyzer (precision 0.5‰) at the Climate Change Institute, University of Maine, and measurements of $\delta^{18}O$ by a Finnigan delta-plus mass spectrometer (precision 0.05‰) at the State Key Laboratory of Cryospheric Science, Chinese Academy of Sciences. All isotopic data are expressed as the relative deviation of the ratio of heavy to light isotope to VSMOW. To test sample reproducibility, 100 samples were selected randomly for duplicate $\delta^{18}O$ measurements. The difference in $\delta^{18}O$ between duplicate analyses of the same sample was on average 0.06‰ (1σ, n = 100). A quadratic error for individual *d* determinations, estimated by the precisions of δD and $\delta^{18}O$, is 0.6‰.

The ice core was annually dated to the year 1534 at the depth of 98 m using seasonal variations of major ions, trace elements and stable isotopes ($\delta^{18}O$ and δD), and the timescale was verified using volcanic horizons (Kaspari et al. 2007). Below 98 m, the core was dated using a flow model because annual layer counting is not possible due to layer thinning (Kaspari et al. 2008). The glacio-chemical records including δD (Kaspari et al. 2007), trace elements (Kaspari et al. 2009), black carbon (Kaspari et al. 2011), snow accumulation rate (Kaspari et al. 2008) and insoluble particles (Xu et al. 2010) of this core have been investigated. However, the *d* record had not been analyzed. Since the 1950s, meteorological observations and atmospheric reanalysis data are available, which assist our investigation into the mechanisms that drive *d* variability.

Hereafter, we focus on the *d* record since 1951 and discuss its variability associated with atmospheric circulation. For the analysias of the atmospheric circulation patterns, meteorological fields including wind vectors, air humidity and geopotential height were calculated based on the National Centers for Environmental Prediction/National Center for Atmospheric Research (NCEP/NCAR) reanalysis data at 2.5° × 2.5° resolution (Kistler et al. 2001). Previous studies indicate that the synoptic-scale variability in temperature and pressure in the Himalayas/Tibetan Plateau can be captured by the NCEP/NCAR reanalysis data (Moore & Semple 2004, Xie et al. 2007), supporting the use of these reanalyses data to depict large scale atmospheric circulation systems around the Himalayan regions.

4. RESULTS AND DISCUSSION

4.1. Seasonal variation

The raw isotopic data ($\delta^{18}O$, δD and *d*) and their annual mean values during the period 1951–2001 are presented in Fig. 2 for all the samples. There is a clear seasonal variation of δD and $\delta^{18}O$, with high

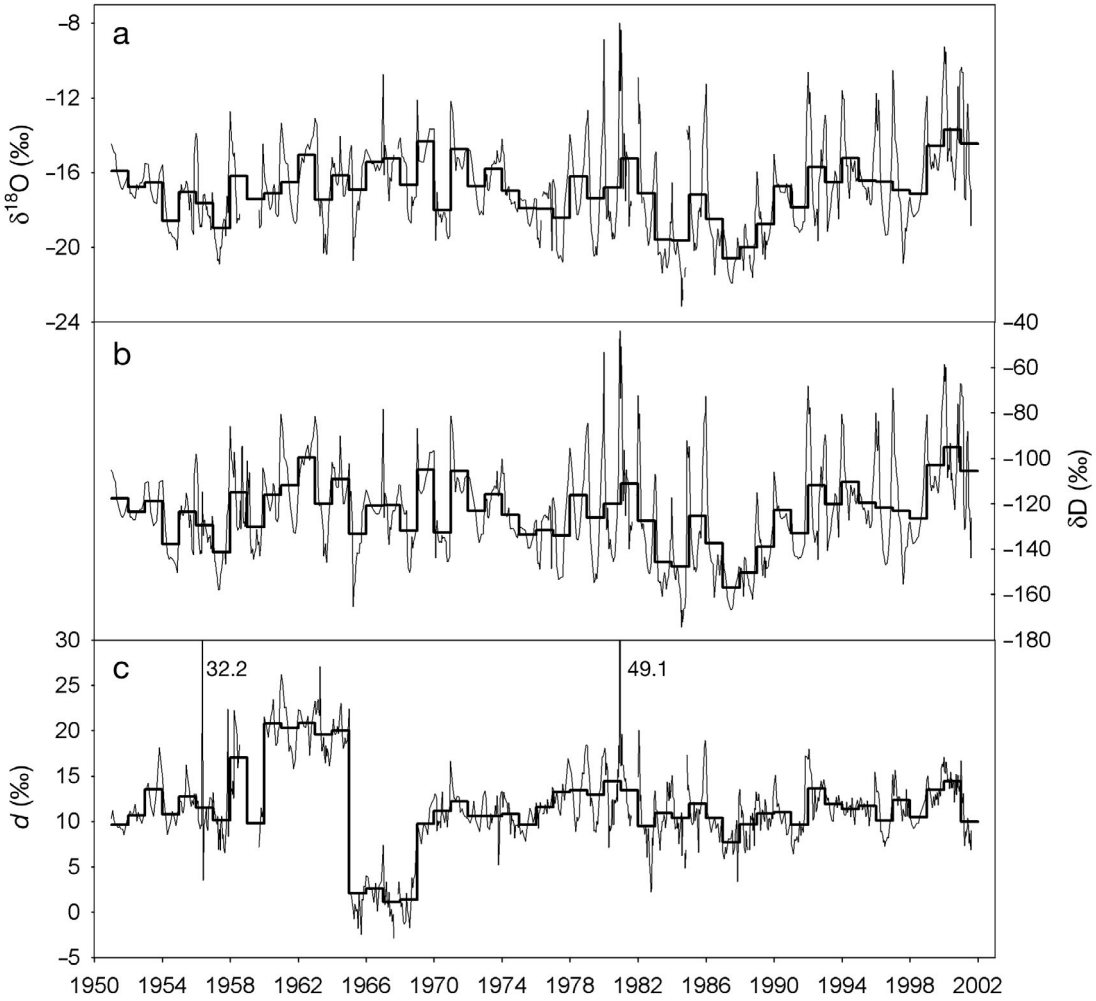

Fig. 2. Raw data of (a) $\delta^{18}O$, (b) δD and (c) deuterium excess (d) measured in the ER ice core for 1951–2001. Thin solid lines: raw data, steps lines: annual mean values. (c) Note: 32.2 and 49.1 are 2 very high d values exceeding the y-axis scale

delta values in winter and low values in summer, and d is in-phase with δD and $\delta^{18}O$ (Fig. 2).

During the wet season in the Himalayan region, the isotopic composition ($\delta^{18}O$ or δD) is primarily controlled by the 'amount effect', which refers to the depletion of heavy water isotopes (more negative $\delta^{18}O$ or δD) during intense summer monsoon rains (Tian et al. 2001, 2003, Lee et al. 2003, Vuille et al. 2005, Pang et al. 2006). Low d values in summer monsoon precipitation are due to limited kinetic evaporation over the Indian Ocean under high surface air humidity conditions. During the dry season, vapor for precipitation generally is derived from continental moisture sources and continental recycling may be important, which would lead to high isotopic values ($\delta^{18}O$, δD and d) in precipitation (Henderson-Sellers et al. 2004, Tian et al. 2007, Froehlich et al. 2008). The contrasting seasonal patterns of d and $\delta^{18}O$ or δD in the ER core are indicative of seasonal shifts of moisture origin linked with changes in large scale atmospheric circulation, as indicated in Section 2.

4.2. Inter-annual variation

4.2.1. 1951–2001

Fig. 2c depicts the inter-annual variation of d. During 1960 to 1964, high d values are observed, with a mean of 20.3 ± 0.6‰, 8.7‰ larger than the 1951–2001 period average (11.6 ± 4.2‰). In contrast, low d values are detected during 1965 to 1968, with a mean value of 1.7 ± 0.7‰, 9.9‰ smaller than the overall average. During the other periods (i.e. 1951–1959 and 1969–2001), no significant multi-annual shift or anomaly can be detected.

It is well known that d in precipitation is largely influenced by conditions prevailing in the oceanic moisture source region from where precipitation originates. Merlivat & Jouzel (1979) developed a theoretical model, in which a kinetic fractionation effect during vapor diffusion from the air-water interface is incorporated, to describe the δD versus $\delta^{18}O$ relationship for global precipitation. The model shows that d primarily depends on the mean relative humidity (h) of the air masses above the ocean surface with a linear relationship of: $d = -58.1h + 57.33$ (Table 1 of Merlivat & Jouzel 1979). The mean surface relative humidity over the tropical Indian Ocean (~10° S–10° N, ~40°–100° E), the Arabian Sea (~10°–20° N, ~50°–70° E) and the Bay of Bengal (~10°–20° N, ~80°–100° E) during the ISM season (June to September) over the period 1951–2001 is 80.1, 74.2 and 83.1%, respectively, calculated from the monthly mean NCEP/NCAR reanalysis data. Assuming that d in the original water vapor is preserved along the air mass trajectory, the simulated d is 10.8, 14.2 and 9.1‰ based on the equation above, respectively. The simulated d value (10.8‰) over the tropical Indian Ocean is in good agreement with the mean value (11.6‰) of d in the ER core during the period 1951–2001. This suggests that the moisture for precipitation at the ER core site is mainly derived from the tropical Indian Ocean.

To verify our conjecture, the mean moisture fluxes during the ISM season (vertically integrated from 1000 to 300 hPa level) along the major trajectories of the ISM, the Arabian trajectory (~10°–20° N, ~50°–70° E), the Bay of Bengal trajectory (~10°–20° N, ~80°–100° E) and the trajectory across the equator from the Southern Indian Ocean (~10° S–10° N, ~40°–60° E), were calculated from the monthly mean NCEP/NCAR reanalysis data, as shown in Fig. 3b-d. The ISM index, defined using the difference of the 850 hPa zonal winds between a southern (~5°–15° N, ~40°–80° E) and a northern (~20°–30° N, ~70°–90° E) region (Wang et al. 2001), and the annual mean of d in the ER core are also presented in Fig. 3. The defined ISM index reflects both the intensity of the tropical westerly monsoon and the lower-tropospheric vorticity anomalies associated with the Indian Monsoon trough. The interannual d in the ER core is positively correlated with the moisture flux along the Arabian Sea trajectory ($r = 0.43$, $n = 51$, $p < 0.01$) and the trajectory across the equator from the Southern Indian Ocean ($r = 0.42$, $n = 51$, $p < 0.01$) during 1951–2001. However, the correlation between the d record and the moisture flux along the Bay of Bengal is not statistically significant. The correlation coefficients improve to 0.74 for the Arabian Sea ($n = 47$, $p <$

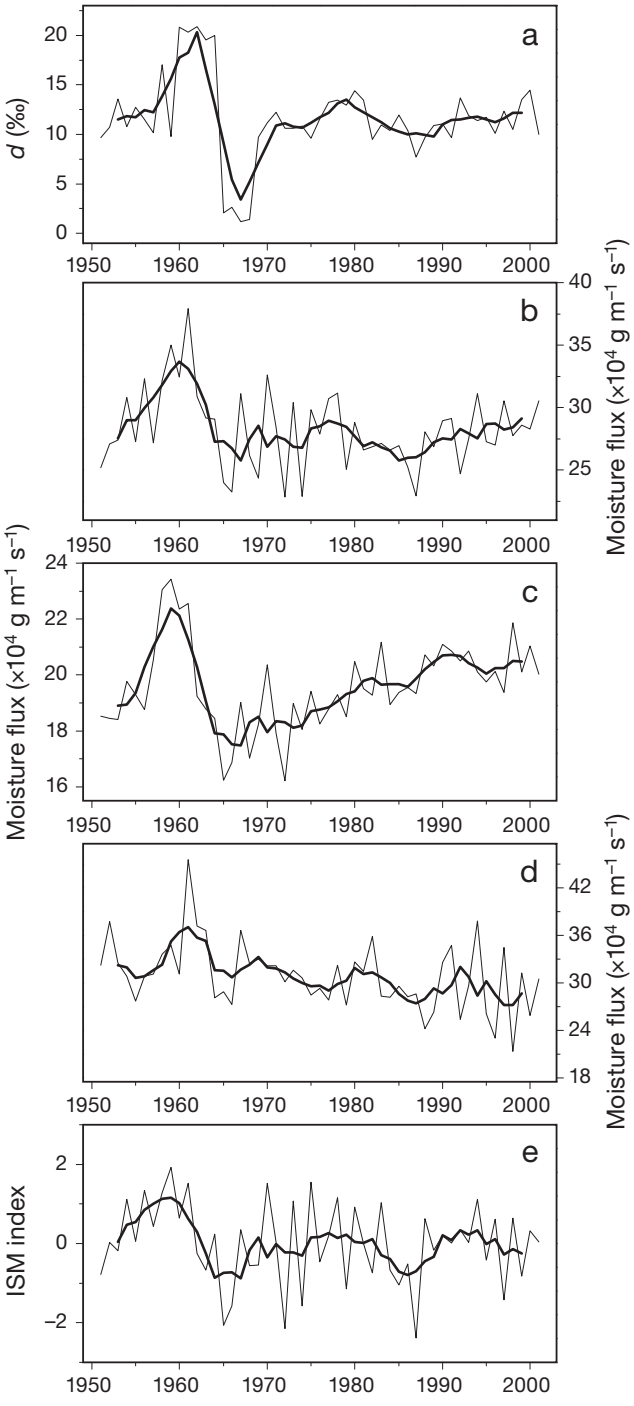

Fig. 3. Variations in annual mean (1951–2001) for (a) deuterium excess (d) in the ER core; (b–d) moisture fluxes during ISM season (vertically integrated from 1000 to 300 hPa level) along its major trajectories; and (e) Indian summer monsoon (ISM) index. Trajectories: (b) Arabian trajectory (~10–20° N, ~50–70° E), (c) across the equator from the Southern Indian Ocean (~10° S–10° N, ~40–60° E), and (d) Bay of Bengal trajectory (~10–20° N, ~80–100° E). Bold lines: 5 yr running means. Note: moisture flux is calculated by multiplying wind vector (m s^{-1}) and specific humidity (g kg^{-1}), and the composite unit is g m^{-1} s^{-1}

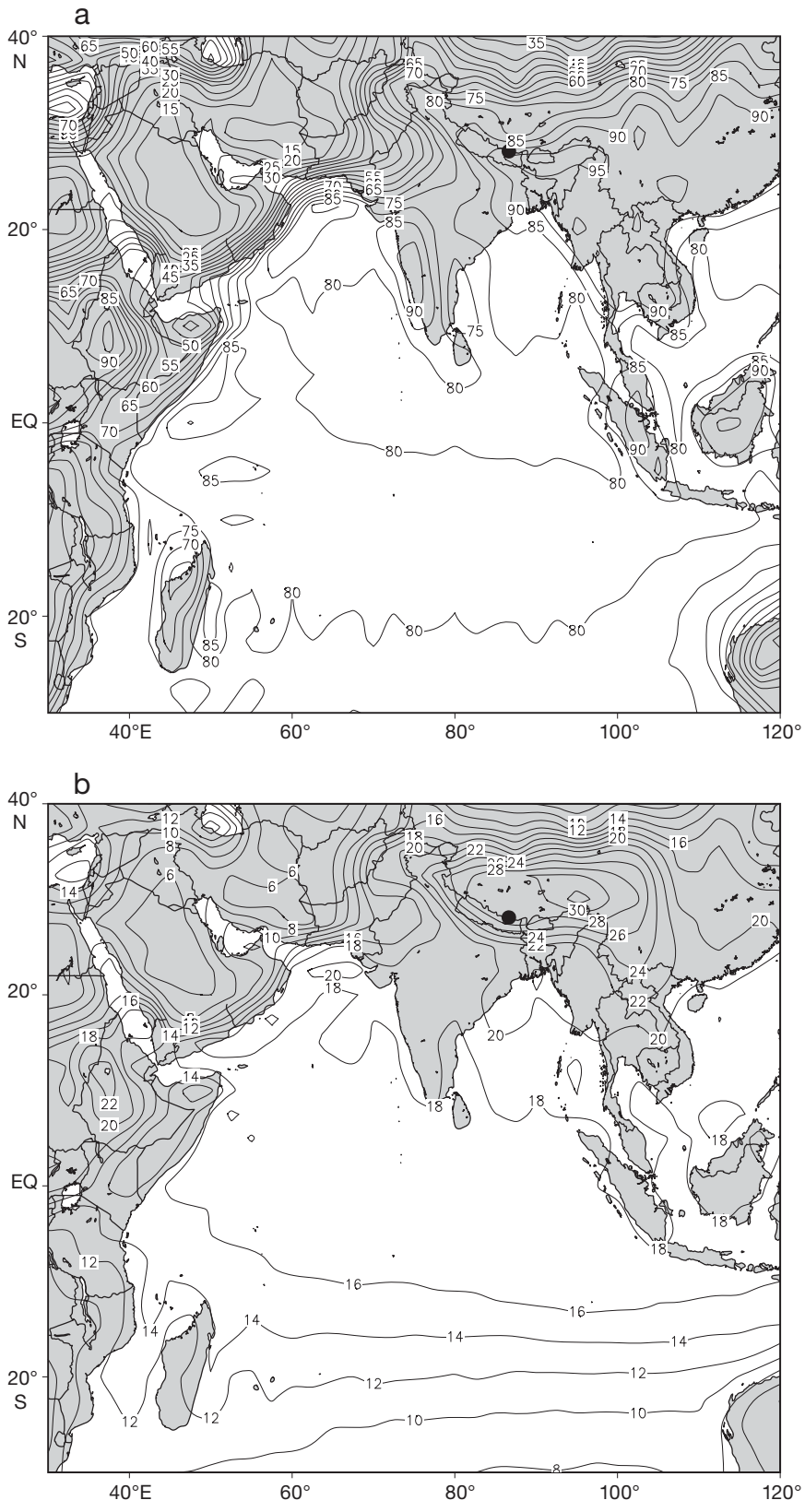

Fig. 4. Distribution of summer (Jun-Sept) mean air humidity in the Indian Ocean over the period 1951–2001. (a) Relative humidity (%) and (b) specific humidity (g kg^{-1}) at 1000 hPa level. (●) ER core site

0.0001) and 0.59 for the Southern Indian Ocean (n = 47, p < 0.0001) when the data are smoothed with the 5 yr running means (Fig. 3). At the same time, the ISM index correlates well with the moisture flux along the Arabian Sea trajectory (r = 0.76, n = 51, p < 0.0001) and the trajectory across the equator (r = 0.65, n = 51, p < 0.0001) during 1951–2001; therefore, the *d* record is correlated positively with the intensity of ISM circulation.

The mean surface air humidity at the surface of the Indian Ocean during the ISM season over the period 1951–2001 was extracted from the monthly mean NCEP/NCAR data (Fig. 4). The spatial distribution clearly shows lower surface air humidity in the tropical Indian Ocean and higher surface air humidity in the north of the Arabian Sea and Bay of Bengal (Fig. 4). When the ISM is stronger, more moisture originating from the tropical Indian Ocean (lower surface air humidity) is transported into the ISM region. On the other hand, more moisture originating from the north of the Arabian Sea and Bay of Bengal (higher surface air humidity) is transported when the ISM is weaker. This explains why the *d* record of the ER core is positively correlated with the intensity of ISM.

However, the abnormally high *d* value (20.3‰) during the period 1960–1964 and the anomalously low value (1.7‰) during 1965–1968 correspond to relative humidity of 63.7 and 95.7%, respectively, based on the equation of Merlivat & Jouzel (1979). It is evident that the anomalous *d* values of 1960 to 1968 cannot be explained reasonably by a change in relative humidity over a fixed ISM moisture source.

4.2.2. 1960–1964

The high *d* value during the period 1960–1964 is indicative of strong kinetic fractionation conditions over the moisture source region. Surface air

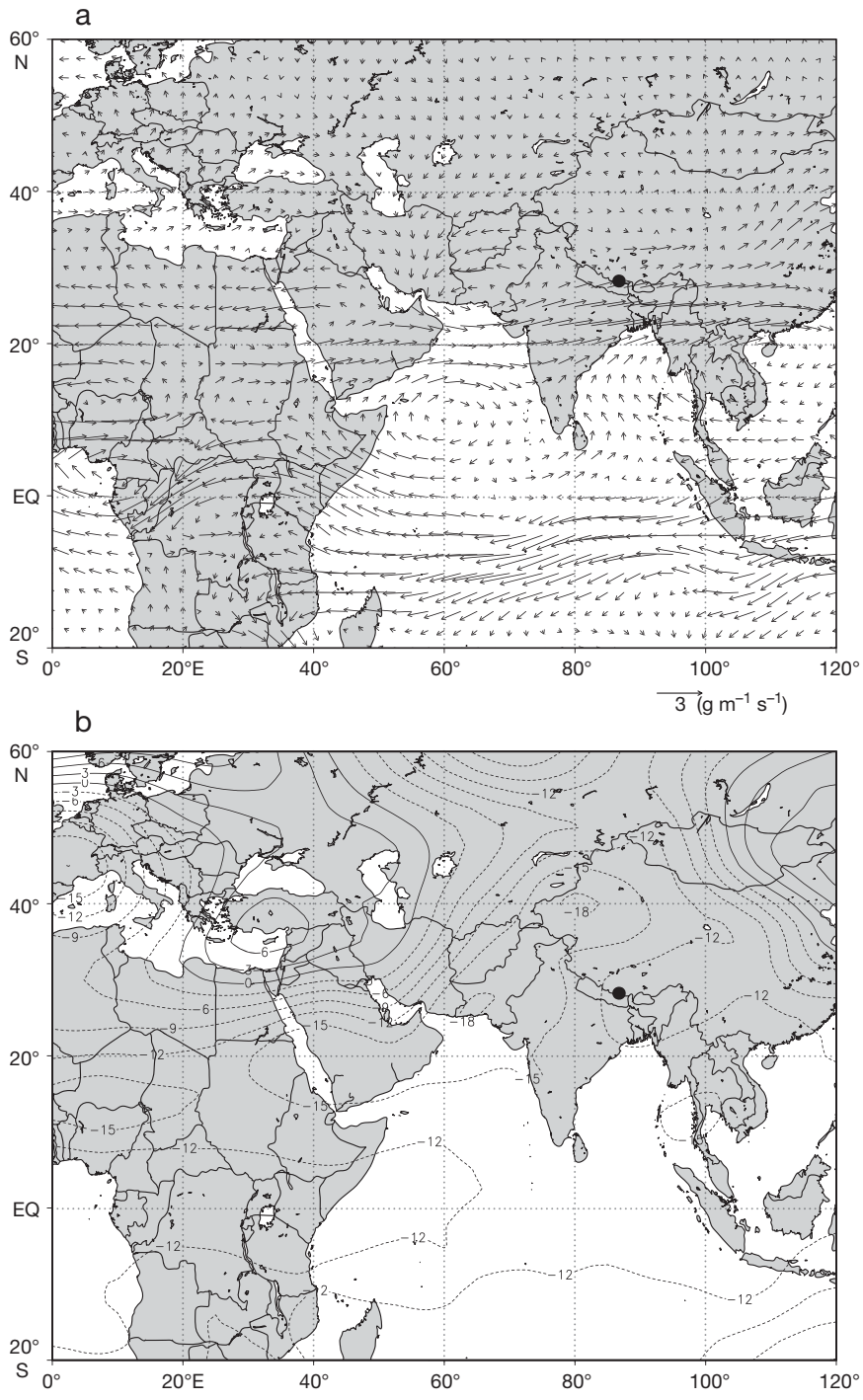

Fig. 5. Anomalies of (a) mean moisture flux (g m^{-2} s^{-1}) at 400 hPa level, and (b) mean geopotential height (m) at 500 hPa level in the dry season (Oct–May of the following year) during 1960–1964 relative to 1951–2001 (1960–1964 minus 1951–2001). Contour interval = 3 m. (●) ER core site

humidity over the ISM moisture source region (the Indian Ocean) during the summer monsoon season is generally high, which limits kinetic effects due to a small difference in surface air humidity at the air-sea interface. Thus, we infer that the high *d* value during the period 1960–1964 is likely related to strong winter westerlies. In Fig. 5, the anomalies of mean moisture flux at 400 hPa level and geopotential height at 500 hPa level during the dry season over the period 1960–1964 relative to 1951–2001 are presented. We can see clearly from Fig. 5a that the moisture transport from the west is strong, suggesting that winter westerlies circulation intensified during the period 1960–1964. The pressure over the 'notch' (30°–32.5° N, 70°–75° E) and most regions of the Tibetan Plateau is low (Fig. 5b), indicating the 'western disturbances' enhanced during this period. The enhancement of 'western disturbances' is expected to increase winter/spring precipitation in the high Himalayas. The air mass for winter/spring snowfall is carried by winter westerlies, with the moisture most likely originating from the northern Atlantic Ocean and augmented by evaporation from the Mediterranean Sea (Med), the Caspian Sea, the Black Sea and Aral Sea (Fig. 1a) (Thompson et al. 2000, Aizen et al. 2005). However, while being transported eastward, these air masses may have experienced strong kinetic fractionation because of dry climatic conditions during the winter/spring season. Previous studies have found that the *d* of moisture from the Mediterranean sea (especially from its eastern part) is very high in winter because of the dry overlying air mass and the strong temperature contrast between the air and the water surface, which results in an isotopic disequilibrium between the moisture in the atmosphere and that of the water body (Gat & Carmi 1970, Rindsberger et al. 1983, Gat et al. 2003, Pfahl & Wernli 2008). As a result, an air mass that experienced strong kinetic fractionation (especially in the continental water bodies) during eastward transport may be one factor that contributes to the high *d* value during the period 1960–1964.

In addition, although the western disturbances dominate the non-monsoon precipitation, moisture from the Northern Indian Ocean (especially the Arabian Sea) still affects winter precipitating water in the Himalayas (Numaguti 1999). In general, air masses associated with winter westerlies are relatively dry. When these dry air masses pass through the Arabian Sea, intense kinetic fractionation is expected to occur and therefore to produce a high *d* value of water vapor. Therefore, non-equilibrium evaporation over the Arabian Sea could be another factor that causes the high *d* of the ER core during the period 1960–1964.

The strong winter westerlies circulation during 1960 to 1964 would transport more moisture to the Himalayan region. Air masses from the west would be trapped by the southern barrier of Himalayas, and it is likely that as air masses are forced repeatedly against this barrier that vapor is transported eastward with several recyclings. Thus, the continental recycling of moisture might be the third factor accounting for the high *d* value during this period.

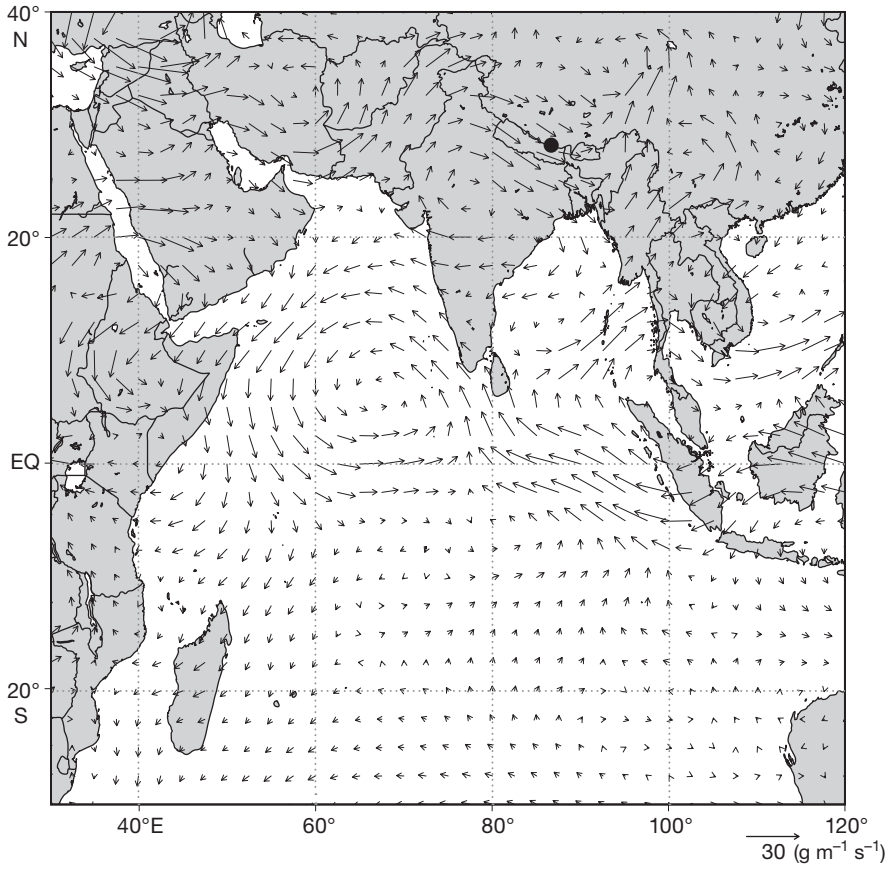

Fig. 6. Anomalies of summer mean moisture flux at 850 hPa level during 1965–1968 relative to 1951–2001 (1965–1968 minus 1951–2001). (●) ER core site

4.2.3. 1965–1968

The ER core *d* values during the period 1965–1968 may imply weak evaporation conditions over the moisture source region. Because the *d* in winter/spring precipitation over the Himalayas is generally high, the low *d* values during 1965 to 1968 are likely related to the ISM or other factors. However, such a sharp decrease of *d* could not be explained by change of surface air humidity over a fixed moisture source region of the ISM. We suggest a major geographical shift of moisture source region of the ER core during the period 1965–1968. In order to verify our conjecture, the difference in summer mean moisture flux at 850 hPa level between the periods 1965–1968 and 1951–2001 is calculated (Fig. 6). A weak ISM circulation is clearly depicted by the moisture flux off the Somali Coast during the period 1965–1968. Although the moisture transport from the tropical Indian Ocean decreases under the condition of weak ISM circulation, moisture transport from the north of the Arabian Sea and Bay of Bengal increases (Fig. 6). This suggests that, when the ISM is weaker, a greater than normal proportion of vapor originating from the proximal source region (i.e. the north of the Arabian Sea and Bay of Bengal) and a less than normal proportion of vapor from the distal source region (i.e. the tropical Indian Ocean) is transported to the ER core site. Thus, it is suggested that the vapor for precipitation at the ER core site during the period 1965–1968 originates mainly from the north of the Arabian Sea and Bay of Bengal, not the tropical Indian Ocean. As shown in Fig. 4, it is clear that the surface air humidity over the north of the Arabian Sea and Bay of Bengal is higher relative to that over the tropical Indian Ocean. Therefore, the geographic migration of moisture source region from the tropical Indian Ocean (with a lower air humidity) to the north of the Arabian Sea and Bay of Bengal (with a higher air humidity) may be a key factor accounting for the low *d* values during the period 1965–1968.

Additionally, the *in-situ* condition at the precipitation site, for example the sublimation of surface snow, could also result in a low *d* in surface

snow (Stichler et al. 2001). The slope of linear regression between $\delta^{18}O$ and δD in snow would decrease if sublimation of snow occurs. Because the fractionation factor (α) of ^{2}H or ^{18}O when the water phase changes from solid to gas are rarely reported, we assume simply that any sublimation of snow at the ER core site occurs at –5°C. Based on the isotope fractionation factor-temperature equations (Clark & Fritz 1997), the calculated value of slope ($s \approx \ln\alpha^{2}H_{ice\text{-}vapour}/\ln\alpha^{18}O_{ice\text{-}vapour}$) is 8.70 at –5°C. The observed values, 8.74 ± 0.13 (n = 78) for the period 1960–1964, and 8.65 ± 0.04 (n = 711) for the periods 1951–1959 and 1969–2001, are in good agreement with the calculated value. However, a significantly lower slope 8.43 ± 0.28 (n = 57) is observed from 1965 to 1968. Under the condition of weak ISM during the period 1965–1968 (Fig. 6), high radiation, low relative humidity and strong winds at the high-altitude glacier sites would occur, which would likely provide good conditions for sublimation. We therefore cannot rule out a contribution of enhanced sublimation to the abnormally low *d* value during the period 1965–1968.

4.3. ISM moisture origin

According to previous studies, there are 2 distinct theories on the origin of ISM moisture: some researchers think that it is mainly formed in the Arabian Sea (Pisharoty 1965, Rao et al. 1981, Murakami et al. 1984), while others suggest that it is from the low-level cross-equatorial moisture flux from the southern Indian Ocean (Pearce & Mohanty 1984, Ramesh Kumar & Schluessel 1998). As shown in Fig. 4, surface air humidity differs for the 2 source regions (higher humidity over the Arabian Sea relative to that over the Southern Indian Ocean). Because *d* in precipitation is correlated negatively with relative humidity over the source region of precipitation (Merlivat & Jouzel 1979), the higher (lower) *d* value at the ER core site likely indicates that greater than normal proportion of vapor from the Southern Indian Ocean (the Arabian Sea) is transported to the ER core site. Good correlations between the *d* record and the moisture fluxes over the Arabian Sea and the sea off the Somalia coast suggest that the Arabian Sea monsoon trajectory and the trans-equatorial trajectory from the southern Indian Ocean are the main channels of moisture to the ER core site. The result seems to mean that the 2 moisture source regions (the Arabian Sea and the southern Indian Ocean) are both sources of precipitation at the ER core site. According to the recent works about global sources of moisture (Gimeno et al. 2010, 2011), both the tropical Indian Ocean and the Arabian Sea could be the moisture sources of the ISM. In this paper, the *d* approach for determining the ISM moisture origin is possibly limited because the *d* record of the ER core is not only controlled by the ISM moisture origin, it may be modified by winter westerlies, moisture transport processes and post-depositional conditions (e.g. snow sublimation). Nevertheless, the close relationship between the *d* record and the ISM moisture flux indicates that the ice core *d* record in the central Himalayas is a good proxy for the ISM.

Although the high Himalayan region is dominated alternately by winter westerlies and the ISM, the seasonal pattern of precipitation is different between the western and eastern Himalayas. For the western Himalayan region, the wintertime precipitation is comparable with the summertime precipitation due to the 'western disturbances'; however, summertime precipitation dominates the annual precipitation in the eastern Himalayas. The good correlation between the *d* record of the ER core and the ISM moisture transport (Fig. 3) suggests that the ER core site is representative of the eastern Himalayan region. However, the 1960–1964 shift of *d* (high values) is likely indicative of an invasion of winter westerlies into the ER core site, suggesting that the ER core site may be the most eastern border of the 'western disturbances'.

Some studies indicate that most annual precipitation in the central Himalayas falls during the summer monsoon season based on the observations from meteorological stations located at low altitudes (Shrestha 2000, Lang & Barros 2004). However, the contribution of winter precipitation at high altitudes in the central Himalayas to annual totals is not well known. Based on the meteorological observation from stations at high altitudes in the central Himalayas, Lang & Barros (2004) found that high elevations (>3000 m MSL) receive up to 40% of their annual precipitation as snowfall during winter. The Nyalam weather station (28° 11′ N, 85° 58′ E, 3810 m above sea level), established in 1967, is located in the central Himalayas, ~100 km west of the ER core site. According to the monthly precipitation data of the station obtained from China Meteorological Administration, the winter/spring precipitation accounts for 53% of its annual precipitation. During 1960–1964, the winter/spring westerlies strengthened; this was associated with (1) an increase in 'western disturbances', (2) transport to the Himalayas of an increased amount of vapour due to strong evapora-

tion from the Arabian Sea, and (3) potential intensification of continental moisture recycling. These factors would probably result in an increase in winter/spring precipitation. As a result, the considerable spring/winter precipitation at the ER core site during the period 1960–1964 probably contributes significantly to the high *d* values during the period; this may be due to different heights of moisture advection by ISM or winter westerlies during 1960–1964.

5. CONCLUSIONS

The first high-resolution *d* record of an ice core drilled from the ER Glacier in the high Himalayas suggests a local response to large-scale atmospheric circulation. The following conclusions can be drawn from the study:

(1) The *d* record of the ER core shows a clear seasonality, with low *d* values in the wet season and high values in the dry season, reflecting seasonal shifts between winter westerlies and the ISM.

(2) The *d* record of the ER core is correlated positively with the ISM moisture flux. The mechanism behind it is the surface air humidity difference between the ISM moisture sources, i.e. relatively high surface air humidity over the north of the Arabian Sea and Bay of Bengal and relatively low surface air humidity over the tropical Indian Ocean (especially the southern Indian Ocean). When more moisture originates from the tropical Indian Ocean, the *d* value of the ER core is higher, whereas the value is lower when more moisture comes from the north of the Arabian Sea and Bay of Bengal. The anomalously low *d* values during 1965–1968 suggest that the moisture for precipitation at the ER core site during this period may originate mainly in the north of Arabian Sea and Bay of Bengal. The close relationship between the *d* record and the ISM moisture flux indicates that the ice core *d* record in the central Himalayas is a good proxy for the ISM.

(3) Although the ER core *d* record is dominated by the ISM circulation, the abnormally high *d* values during 1960–1964 are likely due to the enhancement of winter westerlies. The strong evaporation over the continental water bodies (e.g. the eastern Mediterranean, the Caspian Sea, the Black Sea and the Aral Sea), the intense kinetic evaporation over the Arabian Sea, and the potential intensifying continental moisture recycling, together contribute to the anomalously high *d* values of the ER core during 1960 to 1964.

Acknowledgements. This work was jointly supported in China by National Basic Research program of China (2010CB951401 and 2010CB950301), National Natural Science Foundation of China (40825017, 41171052 and 41176165), Nanjing University and Jiangsu Province; In France, by Centre National de Recherche Scientifique (CNRS) and Service de Coopération Universitaire, Ambassade de France en Chine; and in the United States by the US National Science Foundation (ATM 0139491 and ATM0754644). We are grateful to many scientists, technicians, students and porters for their hard work in the field. Constructive comments by the anonymous referees are much appreciated.

LITERATURE CITED

Aizen VB, Aizen E, Fujita K, Nikitin SA, Kreutz KJ, Takeuchi N (2005) Stable-isotope time series and precipitation origin from firn-core and snow samples, Altai glaciers, Siberia. J Glaciol 51:637–654

Clark I, Fritz P (1997) Environmental isotopes in hydrogeology. CRC press/Lewis Publishers, Boca Raton, FL

Craig H (1961) Isotopic variations in meteoric waters. Science 133:1702–1703

Craig H, Gordon A (1965) Deuterium and oxygen 18 variations in the ocean and the marine atmosphere. In: Tongiogi E (ed) Proc Stable Isotopes in Oceanic Studies and Paleotemperatures, Spoleto, Italy. V. Lishi e F., Pisa

Dansgaard W (1964) Stable isotopes in precipitation. Tellus 16:436–468

Froehlich K, Kralik M, Papesch W, Pank D, Scheifinger H, Stichler W (2008) Deuterium excess in precipitation of Alpine regions: moisture recycling. Isotopes Environ Health Stud 44:61–70

Gat JR, Carmi I (1970) Evolution of the isotopic composition of atmospheric waters in the Mediterranean Sea area. J Geophys Res 75:3039–3048

Gat JR, Klein B, Kushnir Y, Roether W, Wernli H, Yam R, Shemesh A (2003) Isotope composition of air moisture over the Mediterranean Sea: an index of the air-sea interaction pattern. Tellus B 55:953–965

Gimeno L, Drumond A, Nieto R, Trigo RM, Stohl A (2010) On the origin of continental precipitation. Geophys Res Lett 37:L13804, doi:10.1029/2010GL043712

Gimeno L, Nieto R, Drumond A, Durán-Quesada AM, Stohl A, Sodemann H, Trigo RM (2011) A close look at oceanic sources of continental precipitation. Eos Trans AGU 92: 193 doi:10.1029/2011EO230001

Henderson-Sellers A, McGuffie K, Noone D, Irannejad P (2004) Using stable water isotopes to evaluate basin-scale simulations of surface water budgets. J Hydrometeorol 5:805–822

Jouzel J, Stiévenard M, Johnsen SJ, Landais A and others (2007) The GRIP deuterium-excess record. Quat Sci Rev 26:1–17

Kang S, Qin D, Ren J, Zhang D, Qin X (2005) New knowledge of transformation of snow to ice in the East Rongbuk Glacier, northern slope of Mount Qomolangma (Everest). Sci Geogr Sin 25:415–419

Kaspari S, Mayewski P, Kang S, Sneed S and others (2007) Reduction in northward incursions of the South Asian monsoon since 1400 AD inferred from a Mt. Everest ice core. Geophys Res Lett 34:L16701, doi:10.1029/2007GL030440

Kaspari S, Hooke R, Mayewski P, Kang S, Hou S, Qin D (2008) Snow accumulation rate on Mt. Everest: synchroneity with sites across the Tibetan Plateau on 50-100 year timescales. J Glaciol 54:343–352

Kaspari SD, Mayewski PA, Handley M, Osterberg E and others (2009) Recent increases in atmospheric concentrations of Bi, U, Cs, S and Ca from a 350-year Mount Everest ice core record. J Geophys Res 114:D04302. doi:10.1029/2008JD011088

Kaspari SD, Schwikowski M, Gysel M, Flanner MG, Kang S, Hou S, Mayewski PA (2011) Recent increase in black carbon concentrations from a Mt. Everest ice core spanning 1860–2000 AD. Geophys Res Lett 38:L04703. doi:10.1029/2010GL046096

Kistler R, Kalnay E, Collins W, Saha S and others (2001) The NCEP-NCAR 50-year reanalysis: monthly means CD-ROM and documentation. Bull Am Meteorol Soc 82: 247–268

Kreutz KJ, Wake CP, Aizen VB (2003) Seasonal deuterium excess in a Tien Shan ice core: influence of moisture transport and recycling in Central Asia. Geophys Res Lett 30:1922 doi:10.1029/2003GL017896

Lang TJ, Barros AP (2004) Winter storms in the central Himalayas. J Meteorol Soc Jpn 82:829–844

Lee KS, Grundstein AJ, Wenner DB, Choi MS, Woo NC, Lee DH (2003) Climatic controls on the stable isotopic composition of precipitation in Northeast Asia. Clim Res 23: 137–148

Lin Z, Wu X (1990) A preliminary analysis about the tracks of moisture transport on the Qinghai-Xizang Plateau. Geogr Res 9:33–40 (in Chinese with English abstract)

Masson-Delmotte V, Jouzel J, Landais A, Stievenard M and others (2005) GRIP deuterium excess reveals rapid and orbital-scale changes in Greenland moisture origin. Science 309:118–121

Merlivat L, Jouzel J (1979) Global climatic interpretation of the deuterium–oxygen 18 relationship for precipitation. J Geophys Res 84:5029–5033

Moore GWK, Semple JL (2004) High Himalayan meteorology: weather at the South Col of Mount Everest. Geophys Res Lett 31:L18109, doi:10.1029/2004GL020621

Murakami T, Nakazowa T, He T (1984) On the 40-50 day oscillations during the monsoon during the northern hemispheric summer. II. heat and moisture budget. J Meteorol Soc Jpn 62:469–484

Numaguti A (1999) Origin and recycling processes of precipitating water over the Eurasian continent: experiments using an atmospheric general circulation model. J Geophys Res 104(D2):1957–1972

Osterberg EC, Handley MJ, Sneed S, Mayewski PA, Kreutz KJ (2006) Continuous ice core melter system with discrete sampling for major ion, trace element, and stable isotope analyses. Environ Sci Technol 40:3355–3361

Pang H, He Y, Lu A, Zhao J, Ning B, Yuan L, Song B (2006) Synoptic-scale variation of $\delta^{18}O$ in summer monsoon rainfall at Lijiang, China. Chin Sci Bull 51:2897–2904

Pearce RP, Mohanty UC (1984) Onsets of Asian summer monsoon, 1979-1982. J Atmos Sci 41:1620–1639

Pfahl S, Wernli H (2008) Air parcel trajectory analysis of stable isotopes in water vapor in the eastern Mediterranean. J Geophys Res 113:D20104, doi:10.1029/2008JD009839

Pisharoty PR (1965) Evaporation from the Arabian Sea and the Indian southwest monsoon. Proc Symp Meteorol Results, IIOE, Bombay

Ramesh Kumar MR, Schluessel P (1998) Air–sea interaction over the Indian Ocean during the two contrasting monsoon years 1987 and 1988 studied with satellite data. Theor Appl Climatol 60:219–231

Ramirez E, Hoffmann G, Taupin JD, Francou B and others (2003) A new Andean deep ice core from Nevado Illimani (6350 m), Bolivia. Earth Planet Sci Lett 212: 337–350

Rao PLS (2003) Circulation characteristics over the Himalayas during winter season. Meteorol Atmos Phys 83:19–33

Rao GV, Schaub WR Jr, Puetz J (1981) Evaporation and precipitation over the Arabian Sea during several monsoon seasons. Mon Weather Rev 109:364–370

Rindsberger M, Magaritz M, Carmi I, Gilad D (1983) The relation between air mass trajectories and the water isotope composition of rain in the Mediterranean Sea area. J Geophys Res 88:43–46

Schotterer U, Fröhlich K, Gäggeler HW, Sandjordj S, Stichler W (1997) Isotope records from Mongolian and Alpine ice cores as climate indicators. Clim Change 36: 519–530

Shrestha ML (2000) Interannual variation of summer monsoon rainfall over Nepal and its relation to Southern Oscillation Index. Meteorol Atmos Phys 75:21–28

Stichler W, Schotterer U, Froehlich K, Ginot P, Kull C, Gäggeler, Pouyaud B (2001) Influence of sublimation on stable isotope records recovered from high-altitude glaciers in the tropical Andes. J Geophys Res 106(D19): 22613–22620

Thompson LG, Yao T, Mosley-Thompson E, Davis ME, Henderson KA, Lin PN (2000) A high-resolution millennial record of the South Asian Monsoon from Himalayan ice cores. Science 289:1916–1919

Tian L, Masson-Delmotte V, Stievenard M, Yao T, Jouzel J (2001) Tibetan plateau summer monsoon northward extent revealed by measurements of water stable isotopes. J Geophys Res 106(D22):28081–28088

Tian L, Yao T, Schuster PF, White JWC and others (2003) Oxygen-18 concentrations in recent precipitation and ice cores on the Tibetan Plateau. J Geophys Res 108(D9): 4293 doi:10.1029/2002JD002173

Tian L, Yao T, MacClune K, White J and others (2007) Stable isotopic variations in West China: a consideration of moisture sources. J Geophys Res 112(D10):D10112 doi: 10.1029/2006JD007718

Uemura R, Matsui Y, Yoshimura K, Motoyama H, Yoshida N (2008) Evidence of deuterium excess in water vapor as an indicator of ocean surface conditions. J Geophys Res 113: D19114 doi:10.1029/2008JD010209

Vimeux F, Masson V, Delaygue G, Jouzel J, Stievenard M, Petit JR (2001) A 420 000 year long deuterium-excess record measured along the Vostok core: information on past changes in Vostok precipitation origin. J Geophys Res 106:31863–31873

Vimeux F, de Angelis M, Ginot P, Magand O, Casassa G, Pouyaud B, Falourd S, Johnsen S (2008) A promising location in Patagonia for paleoclimate and paleoenvironmental reconstructions revealed by a shallow firn core from Monte San Valentín (Northern Patagonia Icefield, Chile). J Geophys Res 113:D16118 doi:10.1029/2007JD 009502

Vuille M, Werner M, Bradley RS, Keimig F (2005) Stable isotopes in precipitation in the Asian monsoon region.

J Geophys Res 110:D23108. doi:10.1029/2005JD006022
Wang B, Wu R, Lau KM (2001) Interannual variability of the Asian summer monsoon: contrasts between the Indian and the western North Pacific-East Asian monsoons. J Clim 14:4073–4090
Wei K, Gasse F (1999) Oxygen isotopes in lacustrine carbonates of West China revisited: implications for post glacial changes in summer monsoon circulation. Quat Sci Rev 18:1315–1334
Xie A, Ren J, Qin X, Kang S (2007) Reliability of NCEP/NCAR reanalysis data in the Himalayas/Tibetan Plateau. J Geogr Sci 17:421–430
Xu J, Hou S, Qin D, Kaspari S and others (2010) A 108.83-m ice-core record of atmospheric dust deposition at Mt. Qomolangma (Everest), Central Himalaya. Quat Res 73: 33–38

Editorial responsibility: Bryson Bates, Wembley, Australia

Submitted: March 24, 2010; Accepted: February 3, 2012
Proofs received from author(s): April 17, 2012