20. STABLE ISOTOPES THROUGH THE HOLOCENE AS RECORDED IN LOW-LATITUDE, HIGH-ALTITUDE ICE CORES

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1. Introduction

Quantitative use of stable isotopes as proxies for the air temperature at the time of precipitation formation requires establishing relevant transfer functions. Unfortunately, in remote areas where most ice cores are recovered the requisite in situ meteorological observations and contemporaneous precipitation collections are sparse or absent. The processes that determine the values of oxygen and hydrogen isotopic ratios ($\delta^{18}O$ and δD (or ^2H), respectively) in snowfall are complex and so far there is no reasonable model linking the ratios with any single meteorological or hydrologic factor. This is particularly the case for high-elevation, low-latitude glaciers and ice caps, where continental and altitude effects are additional considerations.

Thompson et al. (2003) compared the $\delta^{18}O$ ice histories back to the Last Glacial Stage (LGS) from three low-latitude ice fields in South America and the Tibetan Plateau with similar histories from three polar ice cores. Although these records show large-scale similarities as well as important regional differences, they strongly suggest that, on millennial time scales, changes in $\delta^{18}O$ ice in the tropics represent large-scale climatic variations, including temperature, just as they do in the polar regions. The $\delta^{18}O$ ice records from seven low-latitude, high-altitude glaciers form the basis for this review of how faithfully non-polar ice cores record air temperatures from the end of the LGS through the Holocene.

P.K. Aggarwal, J.R. Gat and K.F.O. Froehlich (eds), Isotopes in the Water Cycle: Past, Present and Future of a Developing Science, 321-339.

The low and mid-latitude ice core recovery sites (Fig. 1) include three from the Tibetan Plateau [Dunde ice cap, (38°N, 5325 m asl), Guliya ice cap, (35°N, 6200 m asl) and Dasuopu Glacier in the Chinese Himalayas (28°N, 7200 m asl)], three from the Andes of South America [Huascarán (Peru, 9°S, 6050 m asl), Quelccaya ice cap (Peru, 14°S, 5670 m asl) and Sajama (Bolivia, 18°S, 6550 m asl) and from Kilimanjaro (Tanzania, 3°S, 5895 m asl)]. The cores from these sites are compared with three polar records, including the Renland (Johnsen et al., 1992) and GISP2 (Grootes et al., 1993) ice cores from Greenland, and the Vostok core from East Antarctica (Jouzel et al., 1987; Petit et al., 1999).

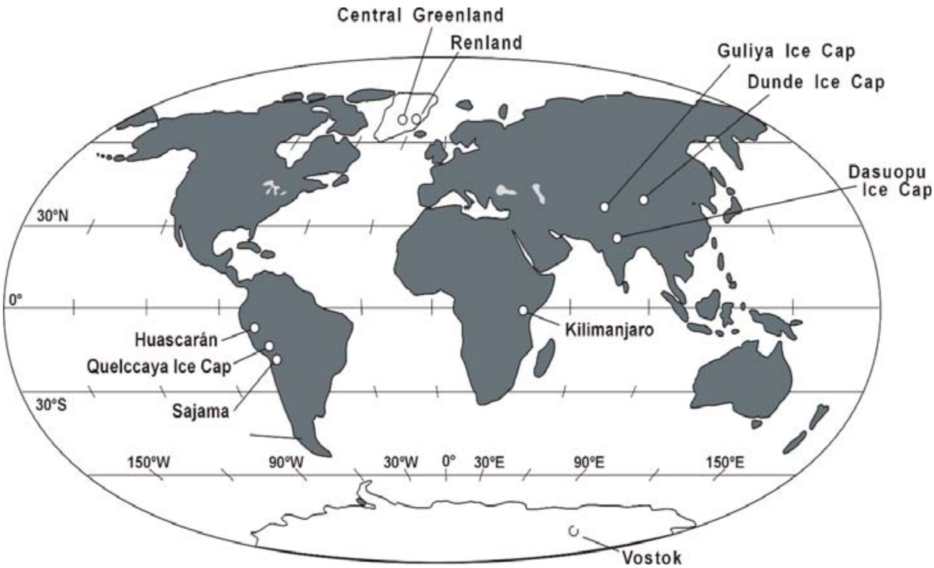


FIG. 1. Locations of ice coring sites discussed in this Paper.

Precipitation is essential for glacier development on high elevation, low and mid-latitude mountain ranges. In these ice core sites the precipitation is highly seasonal. Figure 2 illustrates the modern average monthly precipitation distribution along the tropical Andes of South America, while Fig. 3 illustrates the precipitation across the Tibetan Plateau. Seventy to eighty per cent of the annual precipitation falls in the austral summer (DJF) in the Andes and in the boreal summer (JJA) on the Tibetan Plateau. Along the Andes annual precipitation decreases from north to south while over the Tibetan Plateau both the total annual precipitation and the seasonality decreases abruptly from south to north. This pattern reflects the seasonal cycle in the general circulation of the tropical atmosphere, and more specifically of the Hadley cell with deep convection and associated precipitation on land, warmest sea surface temperatures in the summer hemisphere tropics, and general subsiding motion and dry conditions in the winter hemisphere tropics.

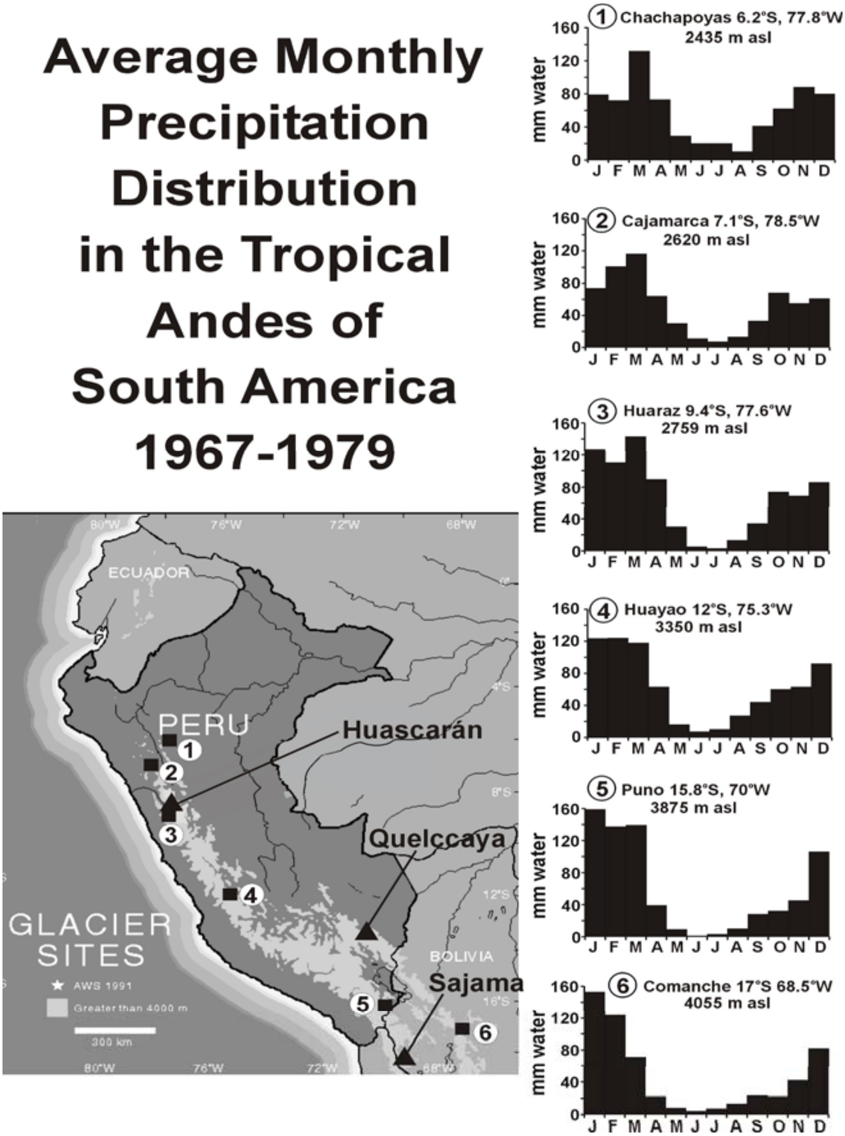


FIG. 2. Meteorological station precipitation records showing the monthly distribution of precipitation under current climate conditions where 70 to 80% of precipitation falls in austral summer (DJF) in the tropical Andes. The locations of the Huascarán, Quelccaya,and Sajama ice core sites are also shown. Holocene ice core $\delta^{18}O$ records and their relation to temperature.

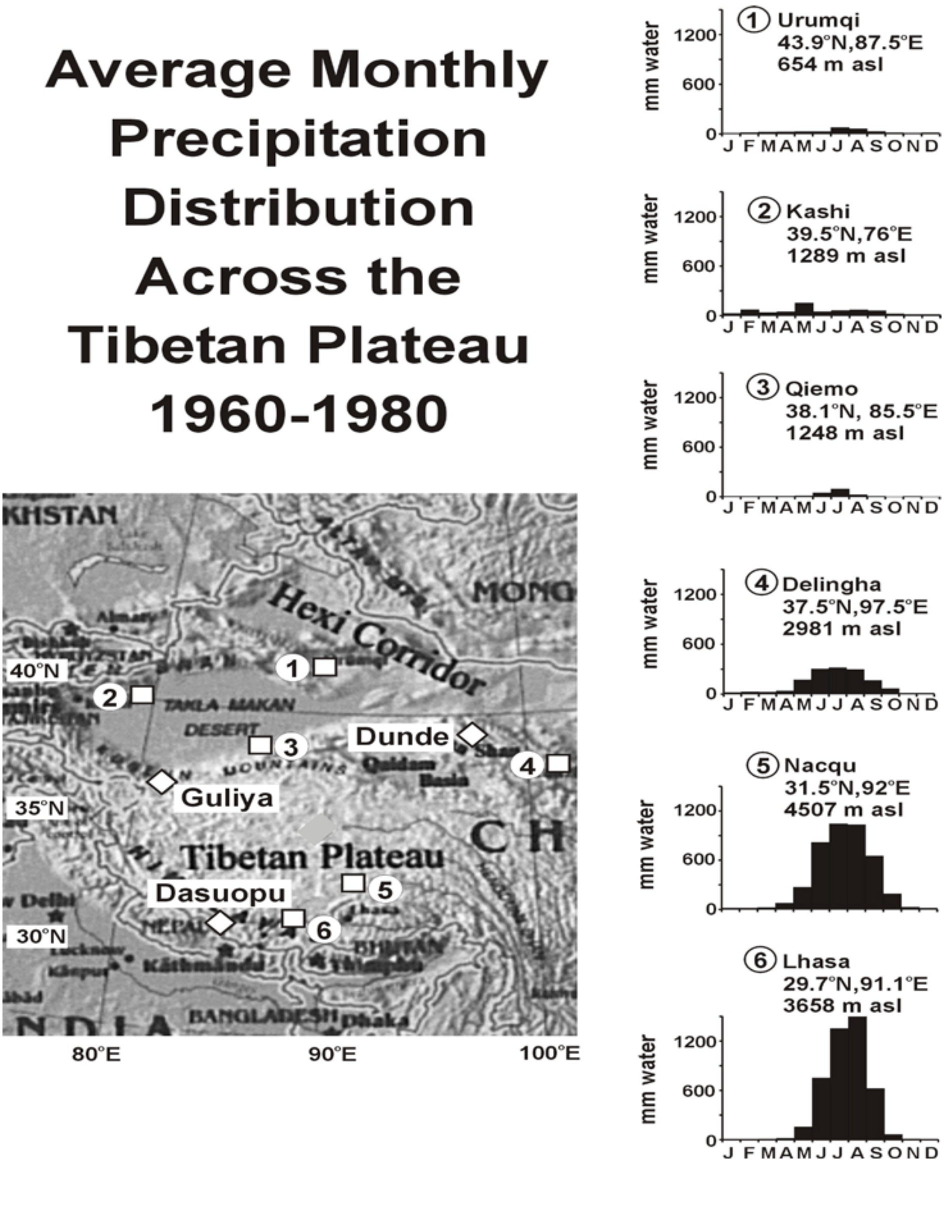


FIG. 3. Meteorological station precipitation records showing the monthly distribution of precipitation under current climate conditions where 70 to 80% of the precipitation falls in the boreal summer (JJA) on the Tibetan Plateau. The locations of the Dundee, Guliya, and Dasuopu ice core sites are also shown.

2. Holocene ice core $\delta^{18}O$ records and their relation to temperature

The $\delta^{18}O$ profiles of the longest low-latitude ice cores (Guliya, Huascarán, Sajama) were presented together in Thompson et al. (2003). The $\delta^{18}O$-proxy temperature difference between the Last Glacial Maximum (LGM) and the Early Holocene (EH) in these cores is similar not only to polar cores, but to other climate histories from other archives such as corals (Guilderson et al., 1994; Beck et al., 1997), noble gases from groundwater (Stute et al., 1995), marine sediment pore fluids (Schrag et al., 1996), snowline depression (Broecker and Denton 1990; Herd and Naeser, 1974; Klein et al., 1995; Osmaston 1965; Porter 1979; Rodbell1992), and pollen studies (Colinvaux et al., 1996). All these records together suggest that the LGM cooling was global.

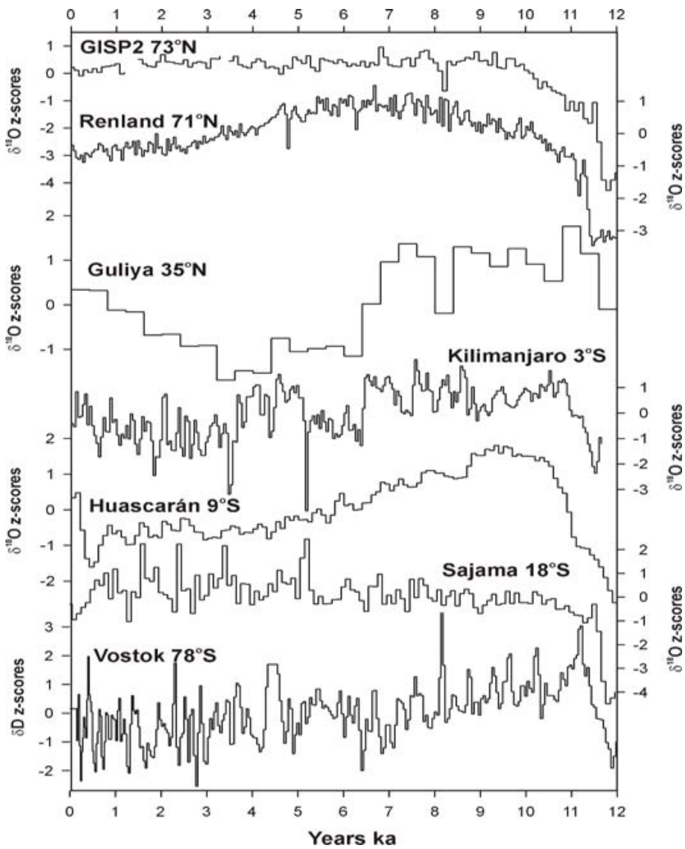


FIG. 4. The $\delta^{18}O$ ice histories for the last 12,000 years from seven cores from the northern polar latitudes through the tropics to the southern polar latitudes. From top to bottom the records are from: GISP2, Greenland (Grootes et al., 1993), Renland, Greenland (Johnsen et al., 1992),Guliya, Tibet (Thompson et al., 1997), Kilimanjaro, Tanzania (Thompson et al., 2002),Huascarán, Peru (Thompson, 2000), Sajama, Bolivia (Thompson et al., 2000), and Vostok, Antarctica (Petit et al., 1999).

Of all ice cores sites throughout the world, Kilimanjaro at 3°S is closest to the equator and thus provides deep tropical records of variations in $\delta^{18}O$, δD, and deuterium (d) excess (Fig. 5). On a global scale, fractionation of oxygen and hydrogen during evaporation and precipitation processes approximates a well-defined relationship, whereby $\delta D = 8\delta O + 10$. This meteoric water line as defined by Craig (1961) is very nearly identical to that illustrated in Fig. 5 for this most equatorial of the world's ice core records. This would suggest that the stable isotopic ratios in these ice fields have not been altered by post-depositional melting or sublimation.

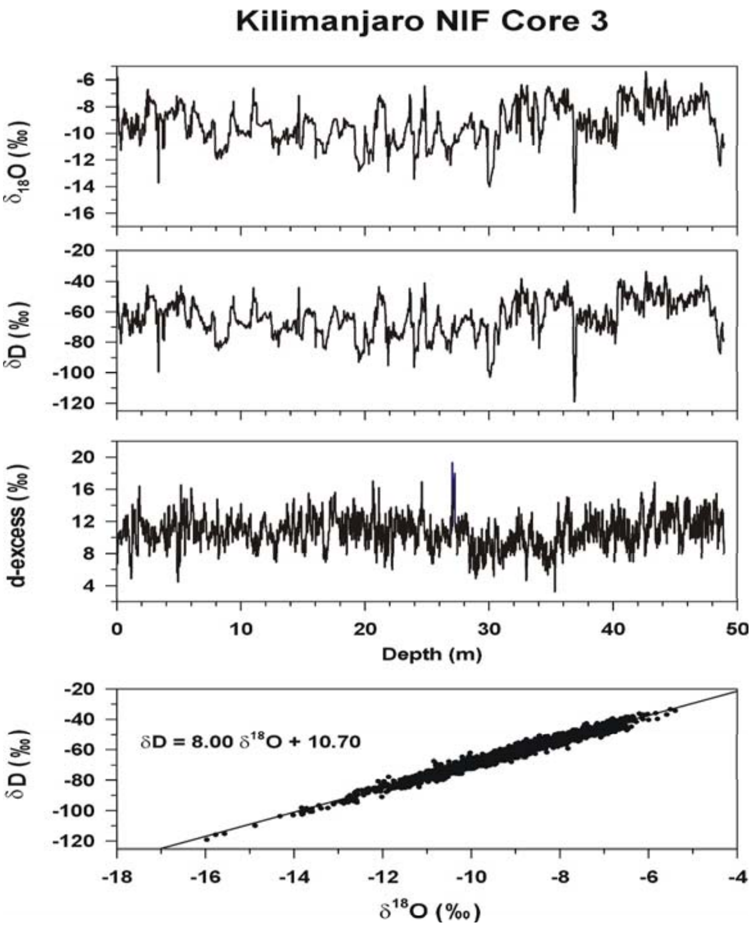


FIG. 5. The $\delta^{18}O$ ice, δD and deuterium excess for Core 3 from the Northern Ice Field (NIF) of Kilimanjaro plotted with depth.

Five of the isotopic records in Fig. 4 (Renland, Guliya, Kilimanjaro, Huascarán, and Vostok) show an Early Holocene "Hypsithermal" warm period (i.e. isotopically enriched) and a Late Holocene "Neoglacial" one (i.e. isotopically depleted). Paradoxically, the two records that do not follow this trend are located relatively close to others that do. The GISP2 ice core record from the summit of Greenland lacks the EH warming, although the drill site is only ~400 km to the east of Renland (which is on the east coast) and at the same latitude (~71°N). The Holocene $\delta^{18}O$ record in Sajama (at 18°S in Bolivia) shows no large-scale variation, which is very different from Huascarán, located 1350 km to the NE along the Andes range.

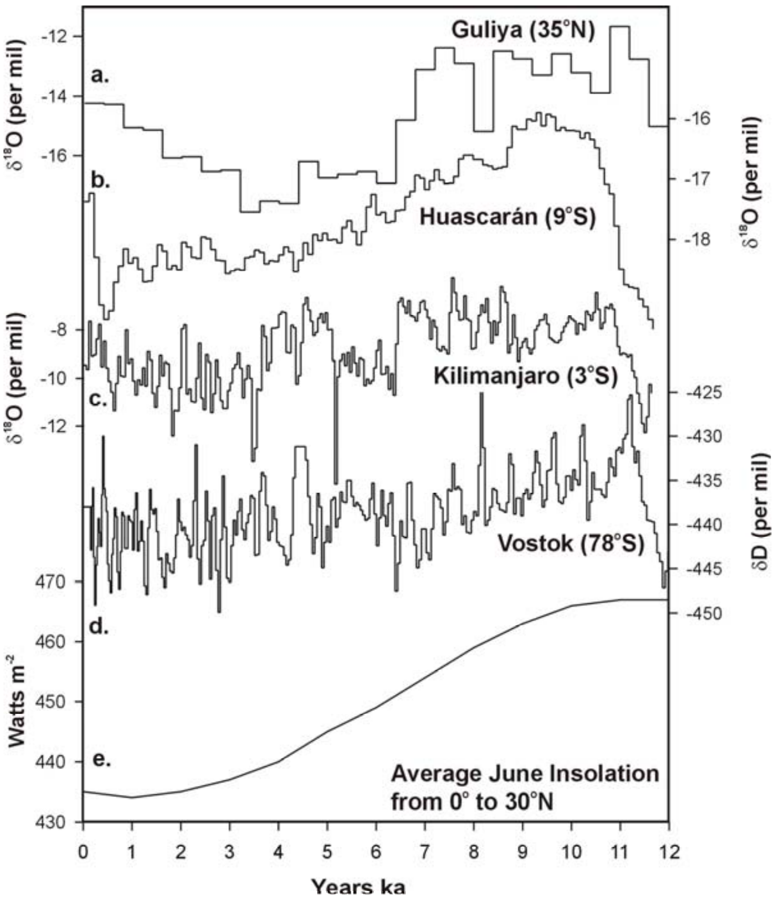


FIG.6. The $\delta^{18}O$ or δD(or ^2H) ice plotted for (a) Renland (Johnsen et al., 1992), (b) Kilimanjaro (Thompson et al., 2002), (c) Huascarán, (Thompson, 2000) and (d) Vostok (Petit et al., 1999) over the last12 thousand years, along with (e) the tropical northern hemisphere (average of 0 to 30°N) insolation changes (Berger and Loutre, 1991).

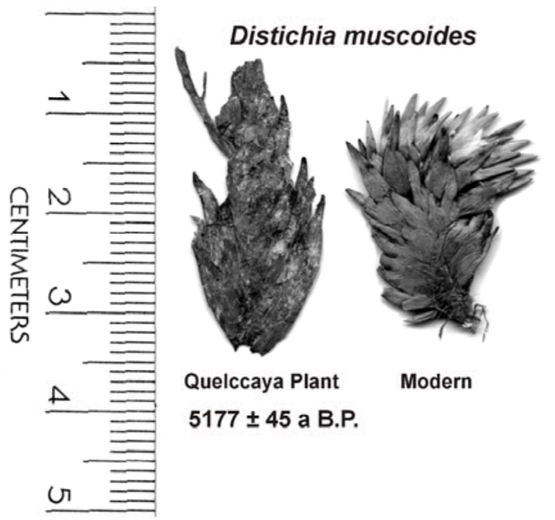
These differences in climate histories between ice cores sites located in close proximity to each other suggest the importance of regional controls on the isotopic record (e.g. the presence of large lakes upwind in the case of Sajama). This is at odds with the idea argued in the literature (e.g. Baker et al., 2001) that tropical $\delta^{18}O$ ice records the regional precipitation, while polar $\delta^{18}O$ ice primarily reflects temperature (Dansgaard, 1964; Dansgaard and Oeschger, 1989). In light of the climatological differences between Huascarán (warm, wet Holocene, Thompson et al., 1995a; 2000a;Thompson, 2000) and Sajama (warm and dry Holocene, Thompson et al., 1998, 2000a; Thompson, 2000) and their similar Holocene $\delta^{18}O$ ice averages, we argue that decadal to century variations in $\delta^{18}O$ ice principally reflect temperature and not precipitation.

The $\delta^{18}O$ or δD records from Renland, Kilimanjaro, Huascarán and Vostok are shown in Fig.6 (a-d). There are similarities between the polar and the tropical isotopic records, suggesting that they are recording the same large-scale variations in the global climate system. All of these isotopic curves mimic the average tropical insulation curve between 0°and 30°N (Fig. 6e), suggesting that the higher temperatures of the EH may have been largely precession-driven. An independent confirmation of the temperature interpretation of the tropical isotopic record was made with the discovery of a plant at the base of the Quelccaya ice cap, located ~800km to the southwest of Huascarán. This plant, shown in Fig. 7, was identified as *Distichia muscoides (Juncaceae*), and was exposed by the receding ice between the austral winter of 2001and its discovery the following year. Interestingly, the altitude where it was found (~5200 m asl) is much higher than the altitude range at which it currently grows (4400 - 4800 m asl). Six samples were recovered and ^{14}C AMS-dated at two laboratories, both of which placed its age at 5177± 45a BP (Fig 7). This date, along with the plant's location, suggests that the climate at this altitude in the Andes was much warmer before ~5.2 ka, which is consistent with the $\delta^{18}O$ records from both Kilimanjaro and Huascarán.

In the longer Huascarán record (Fig. 6c), the early Holocene period was more ^{18}O-enriched than today, indicating warmer conditions which allowed plants to grow at higher elevations than they do under the modern climate. The Kilimanjaro $\delta^{18}O$ record (Fig. 6b) also illustrates this "Hypsithermal" warm period, although the aerosol data also indicate that the climate, while variable, was generally much wetter (Thompson et al., 2002). In fact, the period of ^{18}O enrichment in the Kilimanjaro record corresponds to the well documented "African Humid Period" (e.g. deMenocal et al., 2000). These higher tropical temperatures during the EH may have also determined the mean stable isotopic composition of water vapour in the global atmosphere, which was (and still is) likely dominated by the tropical hydrological system.

3. Controls on isotopic fractionation on precipitation on low-latitude mountains

Two models, which were discussed in some detail in Thompson et al. (2003), have been developed to explain the mechanisms responsible for the $\delta^{18}O$ ice composition found in the Andean ice cores under modern (Holocene) conditions. Grootes et al. (1989) developed a hydrological model of moisture transport which explains the large seasonal difference in $\delta^{18}O$ ice for a region that experiences an annual temperature range of only a few degrees Centigrade. The isotopic composition of ocean water at the source (i.e. the tropical North Atlantic) is modified as the vapour moves across the Amazon Basin and is recycled within thunderstorms when the heavier isotopes are preferentially removed. During the wet season most of the isotopically enriched surface water is transported out of the Amazon Basin by the river system, but in the dry season most of the precipitation that falls in the Amazon is re-evaporated. The ^{18}O-depleted wet season precipitation, which comprises 70 to 80% of the snowfall in the Andes, is further depleted as the air masses are forced to rise over the Andes.



Radiocarbon dates of plants from base of Quelccaya Ice Cap

	^{14}C age	Error (+/-)	Calibrated age (Before 1950 A.D.)	Relative area under probability distribution
Lawrence Livermore National Laboratory				
Sample 1 First run	4470	60	5284-5161 (1σ) 5302-4961 (2σ)	.534 .926
Sample 1 Second run	4525	40	5186-5121 (1σ) 5311-5047 (2σ)	.413 1.000
Sample 2 First run	4530	45	5186-5120 (1σ) 5317-5040 (2σ)	.396 .993
Sample 2 Second run	4465	40	5278-5171 (1σ) 5295-4967 (2σ)	.580 .984
National Ocean Sciences AMS Facility at Woods Hole Oceanographic Institution				
Sample 1	4530	45	5186-5120 (1σ) 5317-5040 (2σ)	.396 .993
Sample 2	4510	40	5188-5119 (1σ) 5307-5040 (2σ)	.404 .988

FIG. 7. The plant Distichia muscoides (5177± 45 a BP) collected at the retreating margin of the Quelccaya ice cap in August of 2002 compared to the modern plant.

An alternative explanation for the $\delta^{18}O$ ice in Andean precipitation includes the important consideration that in tropical mountainous regions snowfall generally originates from thunderstorms with convective cells that extend to great heights such that the mean level of condensation in these storms would be much higher than the condensation level for polar precipitation. More importantly, in the tropics both the geographic location and elevation of the zone of maximum condensation changes from the wet to the dry season. During the height of the wet season, the condensation level is about 2 km higher, where temperatures are cooler and, conversely, in the dry season condensation takes place at a lower, warmer level in the atmosphere.

Thus, the more depleted $\delta^{18}O$ ice values arrive in the wet season snow and could be interpreted as either an 'amount effect' or a temperature signal reflecting changes in the height of the zone of maximum condensation, or some combination of the two (Thompson et al., 2000a). One of the most important questions concerning ice core $\delta^{18}O$ ice records is whether they are a realistic proxy indicator of lower tropospheric temperatures. Observations made by Yao et al. (1996) suggest that on short time scales $\delta^{18}O$ ice trends in precipitation provide a record of temperature trends along the northern part of the Tibetan Plateau. Further to the south, where monsoonal precipitation is more dominant, the $\delta^{18}O$-temperature relationship reverses, so that the most isotopically depleted snow arrives during the summer monsoon (Thompson et al., 2000a). However, as discussed later, intensified atmospheric convection leads to colder condensation temperatures that may be interpreted erroneously as an "amount effect". On annual time scales this $\delta^{18}O$-temperature relationship (e.g., depleted $\delta^{18}O$ ice in the warm, wet season) reflects atmospheric dynamical processes, but over many decades to centuries, atmospheric temperature becomes the dominant process controlling average $\delta^{18}O$ ice.

4. $\delta^{18}O$ history of the tropical Andes over the last millennium

Quelccaya's decadally averaged $\delta^{18}O$ history for the last millennium (Fig. 8(a)) records the "Medieval Warm Period", characterized by more enriched isotopes from A.D. 1100 to 1375, and the "Little Ice Age" characterized by more negative $\delta^{18}O$ lasting until A.D. 1880 (Thompson et al, 1985; 1986). However, the accumulation (An) reconstruction for Quelccaya since 1000 A.D. (Fig. 8(b)) shows little similarity to the isotope record (R2=0.02). The first part of the Little Ice Age, or LIA (1500 to 1720 A.D.) experienced a 30% increase in precipitation, while there was a 20% decrease during the latter part (1720 to 1880 A.D.), but the ^{18}O depletion of 0.9 ‰ occurred consistently through the entire LIA. The lack of similarity between the $\delta^{18}O$ and the An through the last millennium indicates that isotopic ratio of snowfall on Quelccaya is not precipitation-dependent. The decadal averages of $\delta^{18}O$ since 1000 A.D. for the other Andean ice cores, Huascarán and Sajama, were presented individually along with Quelccaya in Thompson et al. (2003).

The composite of these three records is shown in Fig. 9(a), which clearly illustrates the expression in the Andes of the 'Mediaeval Warming' from 1000 to ~1400 A.D., followed by the LIA from~1450 to ~1850 A.D. The recent warming since 1900 A.D. is also very apparent, and the composite in Fig. 9(a) shows that the amount of ^{18}O-enrichment of the last century is the greatest of the last millennium.

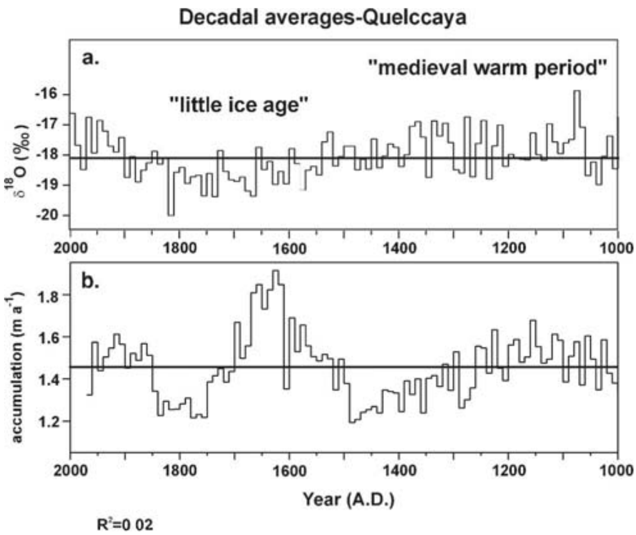


FIG. 8. Decadal averages of (a) $\delta^{18}O$ and (b) accumulation in water equivalent since 1000 A.D. from the Quelccaya ice cap summit core.

5. δ^{18}O history of the Tibetan Plateau over the last millennium

The Dunde ice cap (5325 m asl) on the northeast side of the Tibetan Plateau, the Guliya ice cap(6200 m asl) on the northwest side and Dasuopu Glacier (7200 m asl) on the southern margin form a regional triangular pattern with elevations decreasing from south to north (Fig. 3). The averages of their δ^{18}O ice values since 1000 A.D. (Dasuopu: -20.32‰; Guliya: -14.23 ‰; Dunde: -10.81‰) decrease with increasing altitude (Thompson et al., 2003).

All these glaciers are affected primarily by the Asian Monsoon System in the boreal summer, with minor precipitation in the winter from prevailing westerlies. As was the case for Quelccaya in the Andes, the oxygen isotopic records from these three ice cores bear little resemblance to their corresponding accumulation (An) records. For instance, for the period since 1860 A.D. the R2 between the 5-year averages of δ^{18}O ice and An on Dasuopu is only 0.19, while there is a much stronger statistical relationship between δ^{18}O ice and northern hemisphere temperature anomalies (R2 = 0.37) (Thompson et al., 2000b, their Fig. 6 and Thompson et al., 2003). Visual scrutiny of the δ^{18}Oand An records from the Dunde ice cap since

~1600 A.D. and the Guliya ice cap since 1000 A.D. also show little similarity (Thompson, 1992; Thompson et al., 1995b). These data provide qualitative evidence that temperature, and not the amount of precipitation (or amount effect), is the dominant process controlling δ^{18}O ice over at least this part of the Tibetan Plateau.

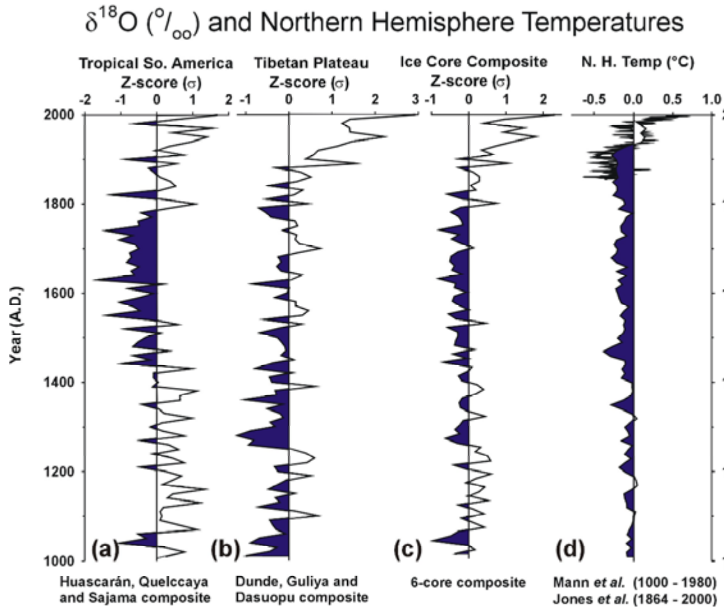


FIG. 9. Regional composites, shown as z-scores, for the last millennium constructed from the decadal averages of δ^{18}O ice from three Tibetan ice cores (a) and three Andean ice cores (b). The composite of all six low latitude cores is shown in (c). The measured (Jones et al., 1999) and reconstructed (Mann et al., 1999) northern hemisphere temperatures are shown in (d) and are plotted as deviations (°C) from their respective 1961-1990 means. Note that the average of δ^{18}O ice for the 1991 to 2000 decade is based on the 1991 to 1997 annual values for the Dasuopu core drilled in 1997 and on the1991-1997 annual values for the Sajama drilled in 1997. The Quelccaya δ^{18}O ice history has been updated to 2000 by drilling new shallow cores.

6. The decadal averages of δ^{18}O ice from the three Tibetan ice cores

The individual δ^{18}O profiles from the Dunde, Guliya, and Dasuopu ice cores display some major differences on both decadal and centennial scales (Thompson et al., 2003). This is not surprising given the diverse regional settings that contribute to differences in precipitation source and post-depositional processes. However, since 1800 all three $^{\delta18}$O ice histories show a consistent trend of enrichment, suggesting that a large spatial-scale warming has affected the region. As with the Andean cores, these three Tibetan cores were combined and their composite is shown in Fig. 9(b). Although the Tibetan Plateau composite shows major differences to the tropical South American composite (Fig. 9(a)), both show an enrichment (warming) trend since 1900 A.D. that is unprecedented in the last millennium. This warming is even more pronounced in Tibet than in the Andes, an observation that is discussed in detail below.

7. 20th century warming

Evidence is accumulating for a strong warming in the tropics in the second half of the 20th century. Although cause and effect are difficult to confirm, it is likely that this warming is the principal driver of the rapid retreat and, in some cases, the disappearance, of ice caps and glaciers at high elevations in the tropics and subtropics. The six cores discussed above provide an opportunity to examine the changes over the last millennium in $\delta^{18}O$ ice at low latitudes in both hemispheres. By their nature, ice cores record fluctuations in the local, regional and larger-scale environment. Thus, they contain local signals that are superimposed upon more regional to global forcings. Unravelling these signals is challenging, given there are only seven low latitude ice cores, including Kilimanjaro, that extend over the last millennium. Integrating the ice core histories with other local proxy records helps in deciphering the local to regional events.

To capture changes at the large spatial scales, we created a tropical composite of the decadal averages of $\delta^{18}O$ ice for the three Tibetan Plateau and the three tropical South American Andes ice cores (Fig. 9(c)). Moreover, since 70 to 80% of the snow in the tropical Andes of South America falls during the wet season (November to April) and on the Tibetan Plateau 70 to 80% of the snow falls in the monsoon season (May to August) (see Figures 2 and 3), combining all six of these records should give a more representative annual, and thus decadal, average $\delta^{18}O$ ice for high elevations in the low latitudes. This composite $\delta^{18}O$ ice record shows enriched $\delta^{18}O$ ice from 1140 to1250 A.D., possibly reflecting the 'Mediaeval Warm Period', and more depleted $\delta^{18}O$ ice from ~1300 to 1850 A.D. correlative with the LIA (Bradley, 2000). However, the most dominant signal in the$\delta^{18}O$ ice composite is the isotopic enrichment in the 20th century.

Figure 9(d) presents the millennial record of decadal temperature variations reconstructed for the last millennium from different types of proxy data primarily from northern hemisphere (NH) locations (Mann et al., 1999). The annual average NH instrumental temperature history from 1864 to 2000 (Jones et al., 1999 and updated from their web site) is also plotted in Fig. 9(d). Both records are presented as deviations from their respective 1961 to 1990 mean values. The similarities between the ice core $\delta^{18}O$ ice composite and the best NH temperature record over the last millennium provide strong evidence that over large distances and decadal and longer time scales, the dominant control on the ice core $\delta^{18}O$ ice record is temperature. The composites illustrate that the 20th century $\delta^{18}O$ ice enrichment is the dominant longer-term (e.g. century-scale) feature common to these regions that are geographically quite separated. The ice core results support meteorological evidence (Jones et al, 1999; Hansen et al., 2001) of a significant 20th century warming, but they have the added value of placing the observations within a longer-term perspective that seems to be signalling a large and unusual warming that is under way at high elevations in the tropics. This is significant as seasonal and annual temperature variations are rather small in the tropics. The average $\delta^{18}O$ ice in Dunde ice deposited since 1950 is enriched by 0.99‰ relative to the millennial mean and similar enrichments on Guliya and Dasuopu are 1.05‰ and 1.84‰, respectively (Thompson et al., 2003). This recent warming is most pronounced at the highest elevation site, Dasuopu, along the southern edge of the Tibetan Plateau. This suggests an amplification of warming at higher elevations in the tropics, as might be expected from the atmospheric thermodynamic considerations discussed above. A recent study (Liu and Chen, 2000) on the Plateau reports a linearly increasing temperature trend of ~0.16°C per decade from 1955 to 1996 and an increasing winter trend of ~0.32°C per decade. Records from 178 stations across the Plateau reveal that the greatest rate of warming (~0.35°C per decade) from 1960 to 1990 occurred at the highest elevation sites. Although these meteorological observations on the Plateau are sparse, they do support the ice core evidence of the enhancement of warming at higher elevations.

Meteorological observations in the tropical Andes are relatively few and of short duration, and there is a similar dearth of data for Tibet. Vuille and Bradley (2000) found that temperature in the tropical Andes has increased by 0.10°C to 0.11°C per decade since 1939. Further, their data indicate that the rate of warming has more than tripled over the last 25 years (0.32°C-0.34°C/decade) and that the last two years of their data series, associated with the 1997/98 El Niño, were the warmest of the last six decades. However, contrary to the findings of Liu and Chen(2000) in Tibet, the rate of warming tends to diminish with increasing elevation. Since 1950 the average $\delta^{18}O$ ice values on Huascarán, the most equatorial site, and Quelccaya have enriched by +1.31‰ and +0.51‰, respectively, over their millennial means. On Sajama, the highest and most southerly site (18°S), the average $\delta^{18}O$ ice in the last 50 years is depleted by 0.20‰ relative to its millennial mean. Clearly, understanding the controls on $\delta^{18}O$ ice in Sajama snowfall warrants additional investigation. A more detailed discussion of some of these controls for recent (since1980) snowfall on Sajama, Quelccaya and Huascarán can be found in Vuille et al. (2003) and Bradley et al. (2003).Other evidence of this high elevation warming is provided by the alpine ice masses that are particularly sensitive to small changes in ambient temperatures as they exist very close to the melting point. Six shallow cores retrieved from the summit of Quelccaya since 1976 document the impact of the recent warming and associated meltwater percolation on the quality of the isotopic signal now preserved in the ice cap's accumulating snow. The sequence of isotopic records from shallow cores taken at the Summit between 1976 and 2002 (Fig. 10) reveals how distinct annual cycles in $\delta^{18}O$ that helped date the upper 600 years in the 1983 cores have not been well preserved at depth since 1991.

Figure 10 also illustrates that the average $\delta^{18}O$ value for the upper 6 to 15 metres of firn has enriched ~3‰ and ~3.5‰ for the upper 6 metres in all cores from 1976 to 2000, contemporaneous with a strong warming in this region (Vuille et al. 2000) that is at least partially responsible for the accelerating rate of glacier retreat. The 2002 summit core shows a better preserved $\delta^{18}O$ and ~2.3‰ depletion since 2000, reflecting a short term fluctuation around the longer-term regional warming trend.

These data suggest that both the oxygen isotopes and the glacier extent are extremely sensitive to large-scale and regional climate changes, but the long-term trend since 1963 remains one of $\delta^{18}O$ enrichment and pronounced, accelerating glacier retreat.

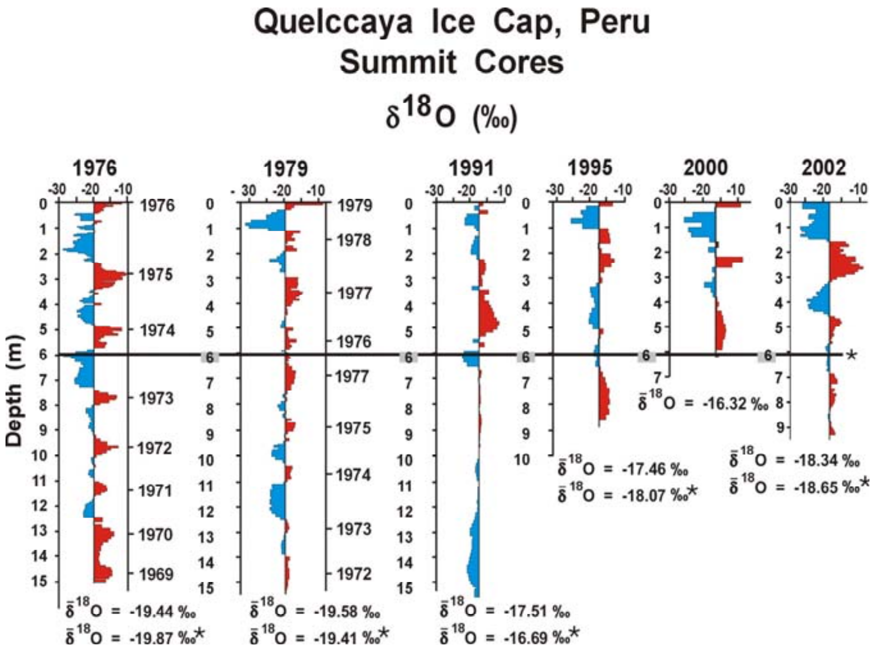


FIG. 10. $\delta^{18}O$ profiles from six shallow cores drilled from the summit of the Quelccaya ice cap from 1976 to 2002, showing the increasing isotopic enrichment and homogenization of the climate signal toward the present. The mean values for each core are shown at the bottom; those marked by an asterisk are the averages from the top of each core to the line at 6 m depth.

The retreat of the Quelccaya ice cap (Peru) is now well documented (Brecher and Thompson, 1993; Thompson et al., 2000a). The observations of the Qori Kalis outlet glacier highlight the sensitivity of these ice fields to ambient air temperatures. These observations document a rapid retreat that has accelerated over the 37-year period from 1963 to 2000. If one takes the 1963 to 1983 rate as a benchmark, Qori Kalis ablated five times faster from 1993 to 1995, eight times faster from 1995 to 1998, and thirty-two times faster from 1998 to 2000. The retreating outlet glacier essentially 'stopped dead in its tracks' from 1991 to 1993 due to the cooling effect of the eruption of Mt. Pinatubo in 1991, but by 1995 it had resumed its rapid retreat. Additional glaciological evidence exists for tropical warming and related ice loss. Hastenrath and Kruss (1992) reported that the total ice cover on Mount Kenya decreased by 40% between 1963 and 1987 and today it continues to diminish.

The Speke glacier in the Ruwenzori Range of Uganda has retreated substantially since it was first observed in 1958 (Kaser and Noggler, 1991). The ice fields on

Kilimanjaro lost 73% of their area between 1912 and 1989 (Hastenrath and Greischar, 1997). During the drilling of Kilimanjaro in 2000, OSU commissioned a new aerial photograph of the summit area, and from that the updated calculation reveals that Kilimanjaro has now lost ~80% of its ice coverage since 1912 (Thompson et al. 2002).

8. Conclusions

Tropical and subtropical ice core records have the potential to provide annual to millennial-scale records of El Niño-Southern Oscillation events and monsoon variability and will continue to provide additional insight to the magnitude and frequency of change in these and other large-scale climate phenomena. The composite low latitude record clearly shows unique changes underway in the 20th century in the low latitudes when viewed from the perspective of the last one-thousand years. The ice cores also contain archives of decadal- to millennial-scale climatic and environmental variability and provide unique insight to both regional and global scale events ranging from the Early Holocene Hypsithermal and the Late Holocene Neoglacial to the Little Ice Age to the recent warming. The data presented here clearly demonstrate that some, if not all, of these unique archives are in imminent danger of being lost if the current warming persists. The urgent need to understand the nature of climate variability in the tropics is illustrated by the geography and demographics of the tropics (± 30° from the equator), which account for 50% of the Earth's surface area, are home to ~70% of the current population of 6.2 billion, produces only ~20% of the world's agricultural goods and accounts for ~80% of the world's births.

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