Forcing of the Asian monsoon on the Tibetan Plateau: Evidence from high-resolution ice core and tropical coral records

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[1] Climatic influences on snow accumulation across the Tibetan Plateau are examined using records of net snow accumulation (A_n) and oxygen isotopic ratios ($\delta^{18}O$) since 1801 from two ice cores from opposite sides of the Plateau. From ~1880 to the 1990s, summer monsoon precipitation has been a significant component of the annual accumulation on the Dasuopu glacier in the Himalayas, but during the latter part of the Little Ice Age (~1810 to ~1880) total A_n was 30% higher than the summer monsoon amounts in northern India. This was possibly the result of increased early winter snowfall as westerly low-pressure systems linked to the North Atlantic pushed farther east along the Himalayas than they normally do today. The decades of high accumulation and the colder temperatures allowed excess snow and ice to persist late into each year, which may have weakened the subsequent Asian summer monsoon. Consequently, precipitation in the northeast Tibetan Plateau where the Dunde ice cap is located may have been affected primarily by Eurasian continental processes rather than tropical meteorology during this time. Since the onset of the recent warming over the last century, the south Asian summer monsoon intensified and influenced summer climate farther to the north and west, and expressions of tropical Pacific and Indian Oceanic/atmospheric processes are noticeable in the Dunde net accumulation record. The Dunde and Dasuopu glaciers, which are located on the northern and southern rims of the Tibetan Plateau, were situated in regions of environmental transition as the Northern Hemisphere climate shifted from a neoglacial to a warming climate mode, which is something to consider when interpreting the longer ice core climate records.

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1. Introduction

[2] The health of the Asian monsoon directly affects 60% of the Earth's population via the availability of water resources for agriculture and industry. Anomalous behavior exhibited by this ocean/atmosphere/land-linked circulation results in hardships from flooding or drought. The essential driver of the monsoon circulation is the thermally induced pressure gradient between land and ocean, and any decrease/increase in that gradient can lead to lower/higher than normal precipitation during a climatically, sociologically and economically critical time of year for Asian populations. Early observational studies of the forcing mechanisms of the monsoons [*Blanford*, 1884; *Walker*, 1910] laid the foundation for later work by *Hahn and Shukla* [1976], *Dickson* [1984], *Rao et al.* [1996], *Kripalani et al.* [1996], and *Wu and Qian* [2003], among others. These, along with model experiments [e.g., *Vernekar et al.*, 1995; *Bamzai and Marx*, 2000], have linked late winter/early spring snow cover and/or depth over various regions of Eurasia inversely with the strength of the subsequent summer monsoon. Large quantities of winter snow that persist into the spring tend to increase the terrestrial albedo and prevent the sensible heating necessary to help drive the monsoon circulation. Most of the observations of Eurasian snow depth and cover are derived from satellite measurements, and thus only extend back to the middle 1960s.

[3] The interaction between tropical ocean processes and the Asian monsoon has recently received much attention, especially through the study of coral records [*Charles et al.*, 1997; *Cole et al.*, 2000]. The El Niño-Southern Oscillation has been linked to monsoon intensity via the zonal Walker circulation [*Walker*, 1910; *Rasmusson and Carpenter*,

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1983], while sea surface temperature and atmospheric dipoles in the Indian Ocean are linked to Pacific Ocean processes on decadal and shorter timescales [*Charles et al.*, 2003]. The ocean basin interactions, and the possible feedbacks between oscillations in land temperatures and SSTs all affect the spatial extent and intensity of the Asian monsoon on interannual to decadal periods.

[4] The presence (or lack thereof) of the connections between all these components of the overall monsoon circulation can be investigated qualitatively further back in time using paleoclimate proxy records of land temperatures, sea surface temperatures (SST), and precipitation. To this end, two centuries of annually resolved ice core data from the northeastern and southern margins of the Tibetan Plateau, along with SST proxy records from western Pacific and western Indian Ocean corals, are analyzed for evidence of monsoon expression and forcing. Paleoclimate records that can demonstrate past climatic relationships between the Asian continent and the tropical oceans and their effects on precipitation, especially prior to the recent acceleration in warming, can greatly aid in reconstructing the long-term history of the summer monsoon and its forcings over the Tibetan Plateau where instrumental records are scarce and of short duration.

2. Data

2.1. Tibetan Plateau Ice Core Records

[5] The Dunde ice cap (38°N 96°E, 5325 meters above sea level, or m asl), located in the Qilian Shan along the northeastern edge of the Tibetan Plateau, was drilled in the summer of 1987 as part of a cooperative program between the Ohio State University's Byrd Polar Research Center (OSU-BPRC) and the Chinese Academy of Science's (CAS) Lanzhou Institute of Glaciology and Geocryology, or LIGG (since absorbed into the Key Laboratory of Ice Core and Cold Regions Environment, or CAREERI). Three cores were drilled to bedrock; core 1 (DnC1, 140 m) was divided between the two institutes, cut into samples, melted and bottled in the field, core 2 (DnC2, 136 m) was taken by LIGG, and OSU-BPRC took possession of core 3 (DnC3, 138 m). The top 55 m of DnC3 was cut into samples, melted, and bottled in the field in keeping with the CAS policy at the time governing sample removal from China, while the lower 83 m of the core was returned frozen to OSU-BPRC. DnC1 was analyzed for δ^{18}O, or the ratios of oxygen-18 (^{18}O) to oxygen-16 (^{16}O), and total dust concentration throughout its length, while DnC3 was analyzed for these parameters plus anion chemistry [*Thompson et al.*, 1989, 1990].

[6] The timescales for the upper sections of DnC1 and DnC3 were constructed using visible dust stratigraphy and verified with seasonal variations in dust and δ^{18}O (Figures 1a and 1b). Each annual layer is bounded by dust layers which are deposited on the glacier during the late winter/early spring as westerly low-pressure systems carry dust from the northwestern China deserts. High aerosol concentrations are generally associated with more negative δ^{18}O values (marked with asterisks); thus the annual layers roughly correspond to calendar years (winter to winter). A beta radioactivity signal, which is a product of the massive Soviet Arctic thermonuclear tests of 1962/1963, appears in the core with the 1963 dust peak, providing a dating calibration. At the top of an ice core where stratigraphic layers are thicker, it is possible to sample continuously at subannual resolution. During the late twentieth century the average net accumulation (A_n) on the Dunde ice cap was ~44 cm water equivalent (w.e.) [*Thompson et al.*, 1989, 1990]. In DnC1, each year since 1801 contains from three to fourteen samples, depending on the layer thickness and sample size. With depth vertical flow causes annual layer thinning, so although the average sample size decreases down-core, eventually so does the number of samples per year until at some point subannual sampling becomes impossible. The annually resolved timescales for DnC1 and DnC3 extend back to 1600 A.D. at 88 and 85 meters depth, respectively.

[7] This study presents only the δ^{18}O time series from DnC1 (Figure 1b), the data from which were produced at the Isotope Laboratory of the Geophysical Institute at the University of Copenhagan. Unfortunately, because of the delay in the purchase and installation of a Finnigan Mat Delta E mass spectrometer at OSU-BPRC, the analysis of the δ^{18}O for DnC3 began almost two years after drilling. Meanwhile, the limited amount of bottled material from the upper part of the ice core was sampled for other parameters (such as dust, anion chemistry, beta radioactivity), and water was distributed among researchers outside OSU for their own work. As a result, the δ^{18}O data set for DnC3 contains numerous gaps, in fact almost 10% of the δ^{18}O data is missing from the upper 55 meters of the core which corresponds to the period between 1808 and 1987.

[8] In 1997, another cooperative program between LIGG and OSU-BPRC culminated in the drilling of the Dasuopu glacier on the north slope of the central Himalayas (28°N, 85°E, 7200 m asl) [*Thompson et al.*, 2000a]. Three ice cores were recovered, the first of which was drilled down-flow from the ice divide but did not reach bedrock. The time series extends back only to 1922 A.D. at 160 m, and below 74 m depth (corresponding to 1945 A.D.) the stratigraphy is severely compromised by ice flow. Cores 2 and 3 (DsC2 and DsC3) were drilled close to the ice divide, and both reached bedrock at 149 m and 168 m, respectively. DsC2 was taken frozen to LIGG, and DsC3 was returned completely frozen and intact to OSU-BPRC. Thus it was possible to sample all of DsC3 in great detail in a controlled environment, which was an advantage over the Dunde analysis. Both DsC2 and DsC3 were analyzed for δ^{18}O in their respective laboratories, but dust concentrations were measured only for DsC3. From the seasonal oscillations of isotopes and dust (as illustrated in Figure 1c for DsC3) time series of annual δ^{18}O variations were produced (Figure 1d). In this ice core, the cold season dust peaks are nearly juxtaposed with the enriched isotopic values. As with the DnC1 record, the beta radioactivity peak of 1963 helped in calibrating the timescale. The cores were annually dated back to 1440 A.D. at 144 meters, with an uncertainty of 3 years [*Thompson et al.*, 2000a].

[9] Annual net accumulation (A_n) records were constructed for both the Dunde and Dasuopu ice cores using a two-parameter steady state flow model that takes into account the rapid thinning of layers near a glacier's flow divide [*Bolzan*, 1985; *Reeh*, 1988; *Thompson et al.*, 1989,

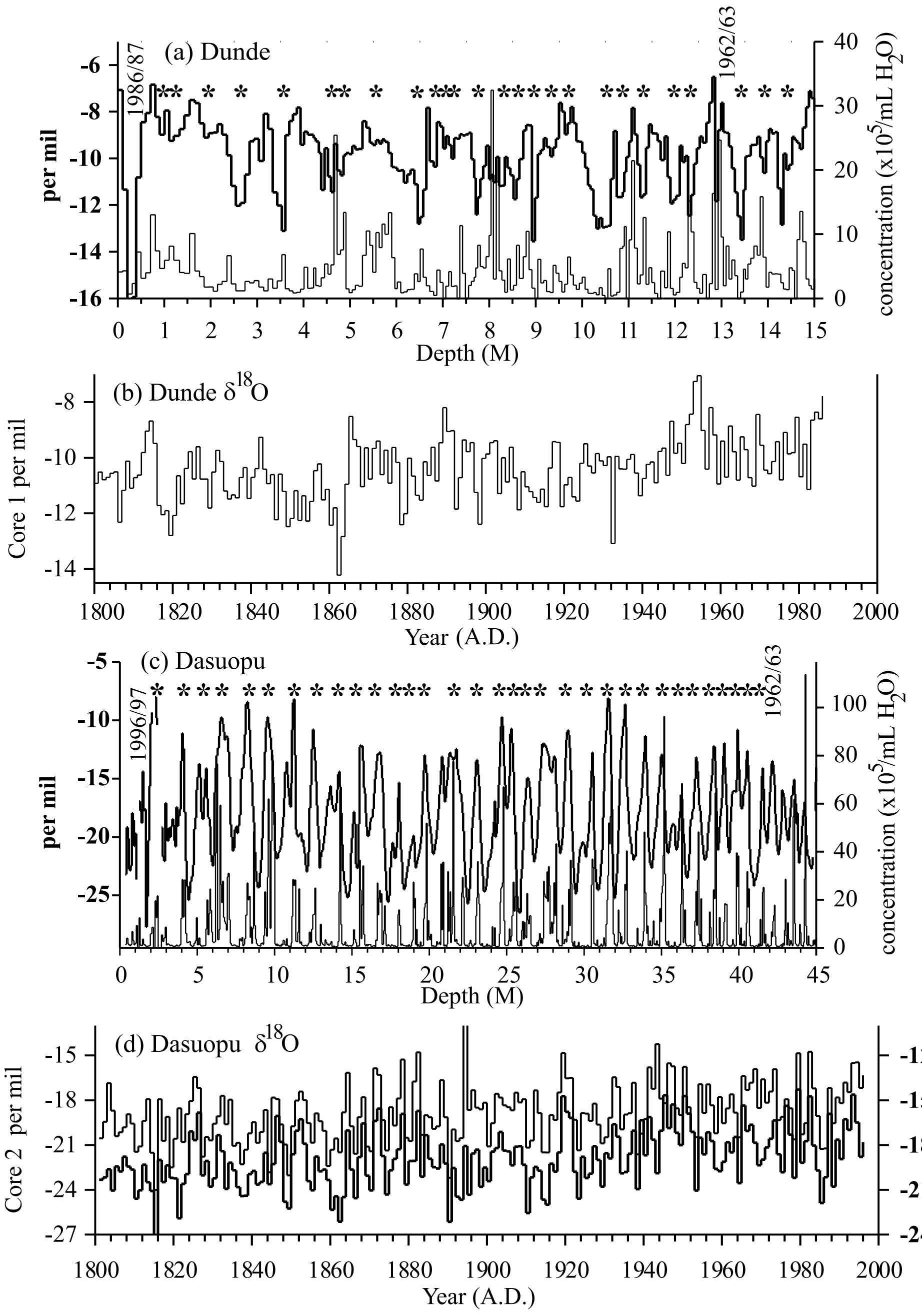

Figure 1. Upper sections of the (a) Dunde core 1 records of dust concentration (light lines) and $\delta^{18}O$ showing seasonal oscillations and the relationships between the parameters as the basis for the annual timescale reconstruction, (b) which is shown from 1801 to 1986. Each winter is depicted in Figure 1a by an asterisk. The beta radioactivity horizon in 1963 from the large 1962/1963 Soviet Arctic thermonuclear tests is also noted. (c) and (d) Same as Figures 1a and 1b for Dasuopu core 3 (heavy line); however, the time series for Dasuopu core 2 (light line) is also shown in Figure 1d.

1990, 2000a]. The reconstructed thickness of each annual layer is determined by the relationship

$$L(z) = b(1 - z/H)^{p+1}, \qquad (1)$$

where H is the ice thickness, b is the average accumulation rate in meters per year (m yr^{-1}), and z is the ice equivalent depth. In a steady state situation, where b is assumed to be constant, the exponent p is uniquely solvable from the depths (z_1 and z_2) of two well-dated horizons at times T_1 and T_2:

$$T_2[(1 - z_1/H)^{-p} - 1] = T_1[(1 - z_2/H)^{-p} - 1]. \qquad (2)$$

For both the Dunde and Dasuopu reconstructions, the upper reference horizon (T_1) was the 1963 A.D. peak in beta activity. The lower time horizon in the Dunde record was 4550 years before 1987 at a depth of 117 m, which was the

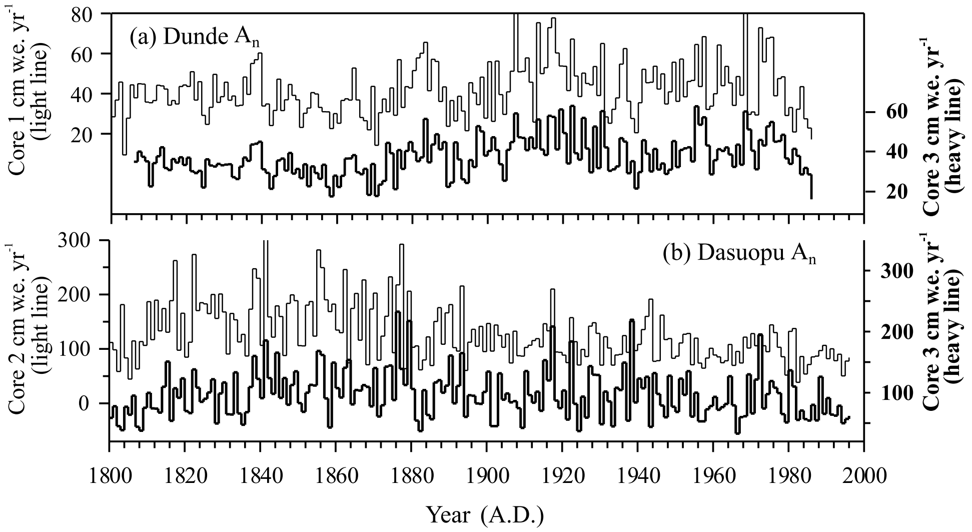

Figure 2. Time series in annual averages for (a) Dunde core 1 net accumulation (A_n) and core 3 A_n; (b) Dasuopu core 2 A_n and core 3 A_n.

limit of the resolution of the annual dust layers in the visible stratigraphy [*Thompson et al.*, 1990], and for Dasuopu it was 1440 A.D., or 557 years before 1997 (at 144 m). The resulting A_n records for the Dunde and Dasuopu ice cores are shown in Figures 2a and 2b, respectively.

2.2. Tropical Coral Records

[10] The $\delta^{18}O$ records from corals have long been recognized as acceptable proxy indicators of sea surface temperature [*Fairbanks and Dodge*, 1979]. The relationship between temperature and $\delta^{18}O$ in corals can be complicated by vital effects and seasonal rainfall variations that might change the isotopic composition of the ocean surface mixed layer [*Cole and Fairbanks*, 1990; *Linsley et al.*, 1994]. Corals that best record SSTs are located in regions where temperature is the major variable. Here, two continuous coral records from the west end of the equatorial Indian Ocean and the western tropical Pacific are used to display SST variations in regions that are important for Asian summer monsoon circulation and other processes with which it interacts. The coral record from the Malindi Marine Park (3°S, 40°E) off the coast of Kenya [*Cole et al.*, 2000] covers the period from 1801 to 1994, and the record from Papua New Guinea (6°S, 148°E) [*Tudhope et al.*, 2001] extends from 1881 to 1993. Both time series are available on the World Data Center for Paleoclimatology/NOAA Paleoclimatology Program web site.

2.3. Reanalysis and Gridded Temperature Data

[11] Time series (1949 to 2003) of surface temperatures from the NCEP/NCAR reanalysis data set [*Kalnay et al.*, 1996] were obtained from the International Research Institute for Climate Prediction Climate Data Library (http://iridl.ldeo.columbia.edu/). The area of interest covers 30°E to 110°E, 60°N to 0°, which encompasses most of Asia from Siberia to the northern Indian Ocean. In addition, longer land temperature records from central Asia were obtained from the NOAA NCDC monthly gridded temperature anomalies derived from the Global Historical Climate Network, or GHCN [*Baker et al.*, 1995], which can be downloaded from http://iridl.ldeo.columbia.edu/SOURCES/.NOAA/.NCDC/.GCPS/.MONTHLY/.GRIDDED/.

3. Concerning the Use of $\delta^{18}O$ in Ice as a Temperature Proxy

[12] Annual values of $\delta^{18}O$ in precipitation show linear relationships with mean annual temperatures at middle and high latitudes [*Craig*, 1961; *Dansgaard*, 1964; *Jouzel et al.*, 1997]; however, the degree to which $\delta^{18}O$ reflects temperature in tropical precipitation is more controversial. Analyses of samples collected from low-latitude regions which experience distinctly different seasonal hydrology demonstrate that there are often inverse correlations between summer $\delta^{18}O$ and the amount of precipitation (the "amount effect"). *Rozanski et al.* [1993] noted that the amount effect has a dominant influence on the depletion of ^{18}O in monsoon precipitation in New Delhi and Hong Kong, located within the Indian and Southeast Asian monsoon regimes, respectively. This relationship is characteristic of tropical precipitation and is documented in ice core records from the Peruvian Andes [*Thompson et al.*, 1986, 1995; *Grootes et al.*, 1989].

[13] An alternative explanation for the lower isotopic values is that a significant portion of the summer precipitation comes from deep convective activity in which fractionation of water vapor occurs during ascent of air parcels within the cold, vertically extensive clouds. In the tropics, the change in mean condensation level (MCL) between wet and dry seasons may result in temperature and humidity conditions that are highly variable. In tropical regions where precipitation results from deep convection during the summer, water vapor condenses at much higher elevations and thus at colder temperatures than during the winter. *Thompson et al.* [2000b] determined the average MCL changes in the atmosphere throughout the year to construct a monthly temperature-isotope relationship in the tropical Huascarán ice core from the Peruvian Andes (9°S, 70°W).

[14] There are several striking differences between the Dunde and Dasuopu dust and $\delta^{18}O$ data, both in seasonal

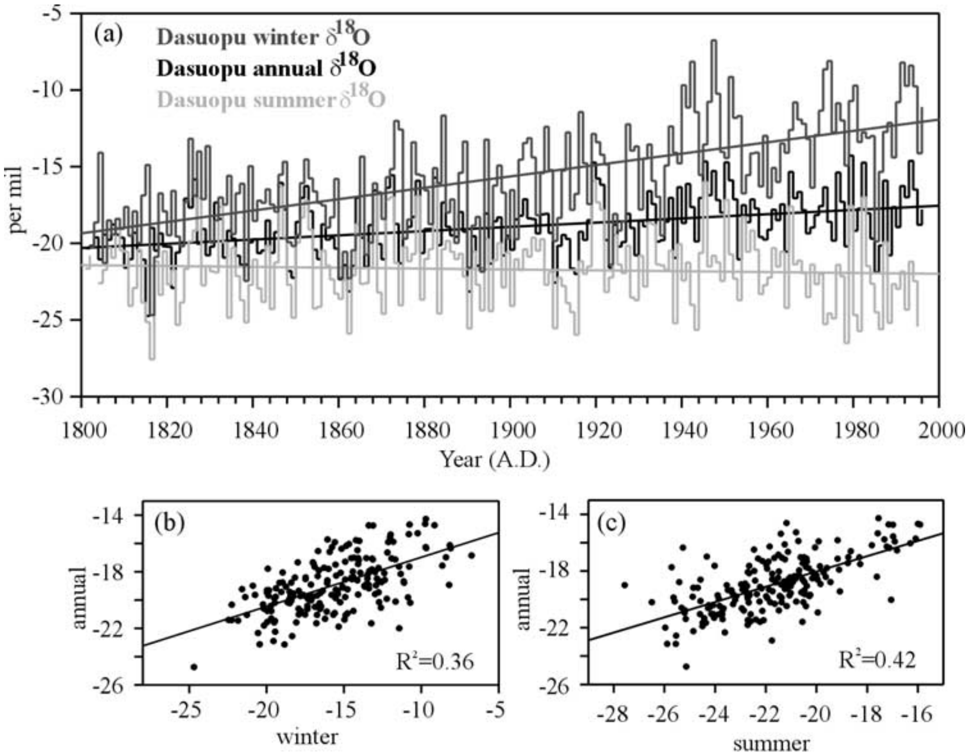

Figure 3. (a) Time series (annual averages) for Dasuopu $\delta^{18}O$, together with the winter seasonal values and summer seasonal values. The regressions between the (b) annual and winter and (c) summer and annual $\delta^{18}O$ time series are at the 99% significance level, N = 195). None of the data are smoothed. See color version of this figure at back of this issue.

timing and in signal-to-noise quality. Because the accumulation rate on Dasuopu is about 2.5 times that on Dunde (currently 100 cm w.e. versus 44 cm w.e., respectively), the wet summer to dry winter amplitudes are much greater and the oscillations vary more smoothly. The monsoon-type seasonal isotopic distribution described above occurs in the southern margin of the Tibetan Plateau where Dasuopu is located [*Thompson et al.*, 2000a; *Dahe et al.*, 2000; *Davis and Thompson*, 2004], with more negative $\delta^{18}O$ (^{18}O depleted) values occurring in the cleaner summer snowfall (Figure 1c). There is a very large vertical relief between the central Himalayas and the plains of Nepal to the south over a short distance, so as the moisture rises, adiabatic cooling and condensation of the water vapor contributes to the isotopic fractionation. Much of the summer snow is derived from the deep monsoon convection at very high-altitude condensation levels, a situation somewhat like that for the Huascarán ice cap described by *Thompson et al.* [2000b]. The winter precipitation, which today comprises a small percentage of the annual total, comes from disturbances that occasionally move eastward from the Hindu Kush-western Himalayan mountain complex [*Rao*, 2003]. The precipitation ultimately originates in the high latitudes, and is delivered to the Himalayas by frontal systems rather than by deep convection. Thus the MCL occurs at a lower elevation in the winter, resulting in less fractionation and more enriched $\delta^{18}O$ values.

[15] As the annual layers thin with depth in an ice core, the amplitudes of the seasonal cycles of $\delta^{18}O$ also decrease. This is due partially to diffusion of the isotopes in situ; however, the decreasing ratio of the number of samples to annual layer thickness down-core is also a critical factor. Normal layer thinning rates in small tropical ice caps accelerate with depth relatively close to the surface; for example, in DsC3 this rapid thinning begins around 60 m, which corresponds to 1940 A.D. As the sample resolution per year decreases, the amplitudes of the seasonal $\delta^{18}O$ variations decrease, and the winter and summer values eventually converge towards the annual averages (Figure 3). The regressions between each detrended seasonal and the detrended annual time series are comparable, as shown in Figures 3b and 3c (0.42 for summer versus annual and 0.36 for winter versus annual). Therefore it is assumed here that the record of annual $\delta^{18}O$ represents the regional temperature variations, which correlates significantly (R = +0.48, significance at the 1% level, 3-year running means) with the Northern Hemisphere June to August (JJA) temperature record of *Jones et al.* [1998]. Thus, whether or not the amount effect plays an intraseasonal role in the isotopic composition of the monsoon-derived snow in the Himalayas, on longer timescales temperatures appear to be more influential [*Thompson*, 2000; *Thompson et al.*, 2000b].

[16] North of the Tanggula Mountains (~33°N), the most enriched (least negative) $\delta^{18}O$ values occur in summer snowfall [*Araguás-Araguás et al.*, 1998; *Tian et al.*, 2001], which is also observed in the Dunde ice core stratigraphy (Figure 1a). Located at 38°N, Dunde is more a temperate than a tropical glacier, and its seasonal isotopic variations more resemble those from polar rather than tropical ice cores. The northeast part of the Plateau, which

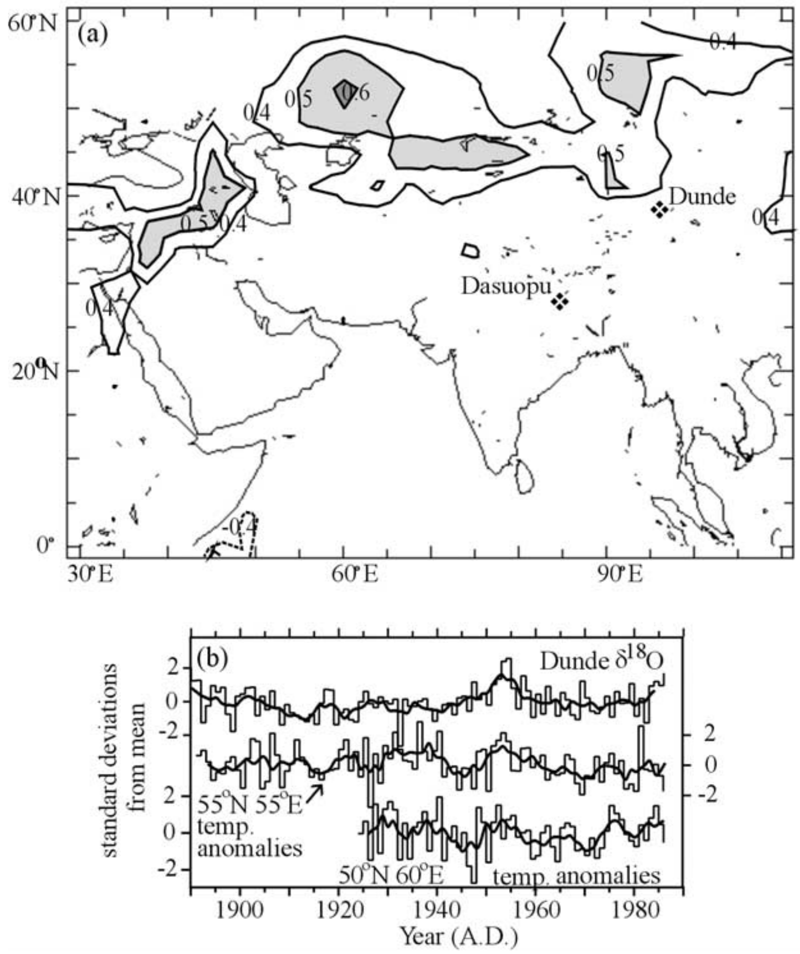

Figure 4. (a) Regions of significant correlation between the Dunde $\delta^{18}O$ record and NCEP/NCAR reanalysis surface temperatures from 1949 to 1986 [*Kalnay et al.*, 1996]. The Dunde time series is treated with three-year running means. Shaded areas indicate correlation coefficients significant at the 99% level. The locations of the Dunde and Dasuopu glaciers are noted. (b) Comparison of the Dunde $\delta^{18}O$ z-scores (standard deviations about the series mean) with extended gridded temperature anomalies from selected coordinates in the high correlation region to the northeast of the Caspian Sea. Heavy lines are five-year running means.

lies on the edge of the Asian monsoon regime, receives summer moisture not only from the south but also from mid-latitude westerlies and from the north [*Yatagai and Yasunari*, 1995]. In the past the Asian monsoon boundary has shifted in response to warming [*Liu et al.*, 1998]; for example, during the mid-Holocene "climatic optimum" its influence may have extended well to the north beyond its present limits. Any monsoon-derived moisture that the Qilian Shan receives has been recycled over thousands of kilometers over land. Since the winter and summer precipitation form at equivalent MCLs, it is most likely that surface temperature has the strongest control over the isotopic fractionation.

[17] There are several lines of evidence supporting the link between $\delta^{18}O$ values in precipitation and temperature in the northern part of the Tibetan Plateau. Significant relationships exist between monthly averages of $\delta^{18}O$ measured in precipitation collected at Delingha (150 km from Dunde) and air temperature [*Yao et al.*, 1996]. On decadal timescales, $\delta^{18}O$ in a core from the Malan ice cap (36°N, 90°E), which is also located close to Dunde, has demonstrated significant correlations with 500 hPa temperatures at stations across the northern Plateau over a 30-year period in the late twentieth century [*Wang et al.*, 2003]. In order to show the spatial patterns of the correlations between Eurasian surface temperature and the recent part of the Dunde oxygen isotope record, the annual averages of $\delta^{18}O$ from 1949 to 1986, smoothed with three-year running means (3 yrm), are correlated with unsmoothed mean June–July–August (JJA) surface temperatures from the NCEP/NCAR reanalysis data set [*Kalnay et al.*, 1996] over Asia (Figure 4a). The contours outlining regions of significant correlation coefficients (R) are shown, and those that are significant at the 99% level ($R > +0.5$, $N = 37$) are shaded. These areas extend throughout Asia north of 40°, and are even located close to Dunde, but the region of highest correlation ($R > +0.6$) is in the Kirgiz Steppe to the northeast of the Caspian Sea.

[18] The reanalysis data only extends back to 1949, but the gridded temperature anomalies that are derived from the GHCN provide continuous reconstructed records based on

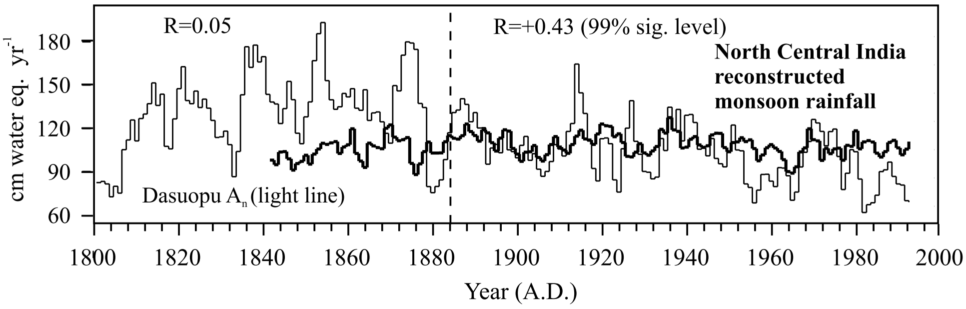

Figure 5. Comparison between the Dasuopu A_n record since 1801 and the north central India (NCI) reconstructed monsoon rainfall record from *Sontakke and Singh* [1996]. Correlation coefficients are shown for the periods 1840 to 1883 and 1884 to 1986. Both time series are smoothed with three-year running means. The standard error of the estimated regression for NCI values ranges from 119 mm at 1847 to 19 mm at 1871.

instrumental measurements back to the end of the nineteenth century in a few places in Asia. Fortunately, two of those grid points lie in the region of highest surface temperature and $\delta^{18}O$ correlation, at 50° to 55°N and 55° to 60°E. Figure 4b illustrates that the average June to August (JJA) temperature anomalies for these coordinates and the annual averages of the Dunde $\delta^{18}O$ show similar trends over decadal timescales.

4. Dasuopu A_n Record Since the Little Ice Age

[19] Recent studies indicate that the glaciers in the central Himalayas currently receive most of their annual snowfall (up to 70%) during the summer monsoon season from the Arabian Sea via north central and northeast India [*Thompson et al.*, 2000a; *He et al.*, 2003]. In fact, the Dasuopu A_n record (DsC2 and DsC3 combined) from ~1880 to ~1995 broadly resembles the reconstructed summer monsoon rainfall record from north central India (NCI) [*Sontakke and Singh*, 1996], with accumulation in both averaging about 100 cm yr^{-1} of water during this period (Figure 5). Since 1950, the two records have diverged slightly as Dasuopu A_n has slowly decreased, possibly due to recent glacier wastage that is affecting much of the Himalayas and the Tibetan Plateau [*Kadota et al.*, 1997; *Dahe et al.*, 2000; *Su and Shi*, 2002; *Liu et al.*, 2003]. During the latter part of the Little Ice Age (LIA) before 1880, the Dasuopu A_n averaged ~30% higher than the NCI rainfall, and in fact higher than any of the regional or all India summer monsoon indices compiled by *Sontakke and Singh* [1996]. The excess snow during this latter part of the LIA may have been the result of either higher summer precipitation that somehow was not experienced in the monsoon regions to the south, or by higher accumulation rates during the winter.

[20] The winter westerly storms that originate as cutoff lows near the Caspian Sea [*Rao*, 2003] currently deliver most of the annual snowfall to the western Himalayas. These "western disturbances" travel through the intersection, or "notch" between the western Himalayas and the Hindu Kush Mountains, boxed by the coordinates 30 to 32.5°N, 70° to 75°E [*Lang and Barros*, 2004]. The disturbances are low-pressure systems, characterized by the lowering of the 500 hPa geopotential height (GPH) in the region, that contribute to orographically forced snowfall further to the east in the central Himalayas. A weak but significant correlation between the low-pressure systems and the Polar/Eurasian teleconnection suggests that the atmospheric conditions in the circumpolar region play a role in the intensity and frequency of the winter storm activity in these mountains.

[21] Significant correlations between the summer North Atlantic Oscillation (NAO) and the concentration of sodium chloride in a shallow Dasuopu ice core suggest a direct or indirect teleconnection between atmospheric processes in the North Atlantic and the meteorology in the Asian monsoon regime, including the Himalayas [*Wang et al.*, 2002]. Following this hypothesis, it may be reasonable to infer such a connection, indirect as it may be, between the North Atlantic and the central Himalayas via the western Himalayan "notch" in the winter, when the effects of the NAO are more significant across Eurasia. *Phillips et al.* [2000] in fact noted a sensitivity of Himalayan glaciation to North Atlantic circulation changes on millennial or shorter timescales. Just as there is a weak but significant relationship between the notch GPH and the Polar/Eurasian teleconnection [*Lang and Barros*, 2004], there is a similar correlation (when both time series are smoothed with three-year running means) between the former and the NAO index (NAOI) during the late autumn and winter months (November to March) from 1949 to 2003 (Figure 6a).

[22] After comparing the extended NAOI time series [*Jones et al.*, 1997] during various combinations of winter months with the Dasuopu A_n, the closest match was found unexpectedly using the November and December combined indices, but not with indices later in the winter (Figure 6b). However, the similarities exist only before the early 1880s, when the records show a significant correlation ($R = -0.32$, 95% significance level). Around 1880, as the LIA ended and the Northern Hemisphere temperatures warmed, the Icelandic Low strengthened and within 10 years the NAOI shifted from a strong negative to a positive mode, contemporaneously with a sharp drop in the A_n.

[23] On visual inspection of the Dasuopu A_n and NAO records, it is obvious that strong oscillations are present that change in frequency and amplitude between the LIA and the post-LIA periods. Spectral analysis using the Blackman-Tukey method (with a Bartlett window) was performed on these subseries. Prior to 1880, the spectral curves of the November–December NAOI and Dasuopu A_n are quite similar, with both series showing significant signals at

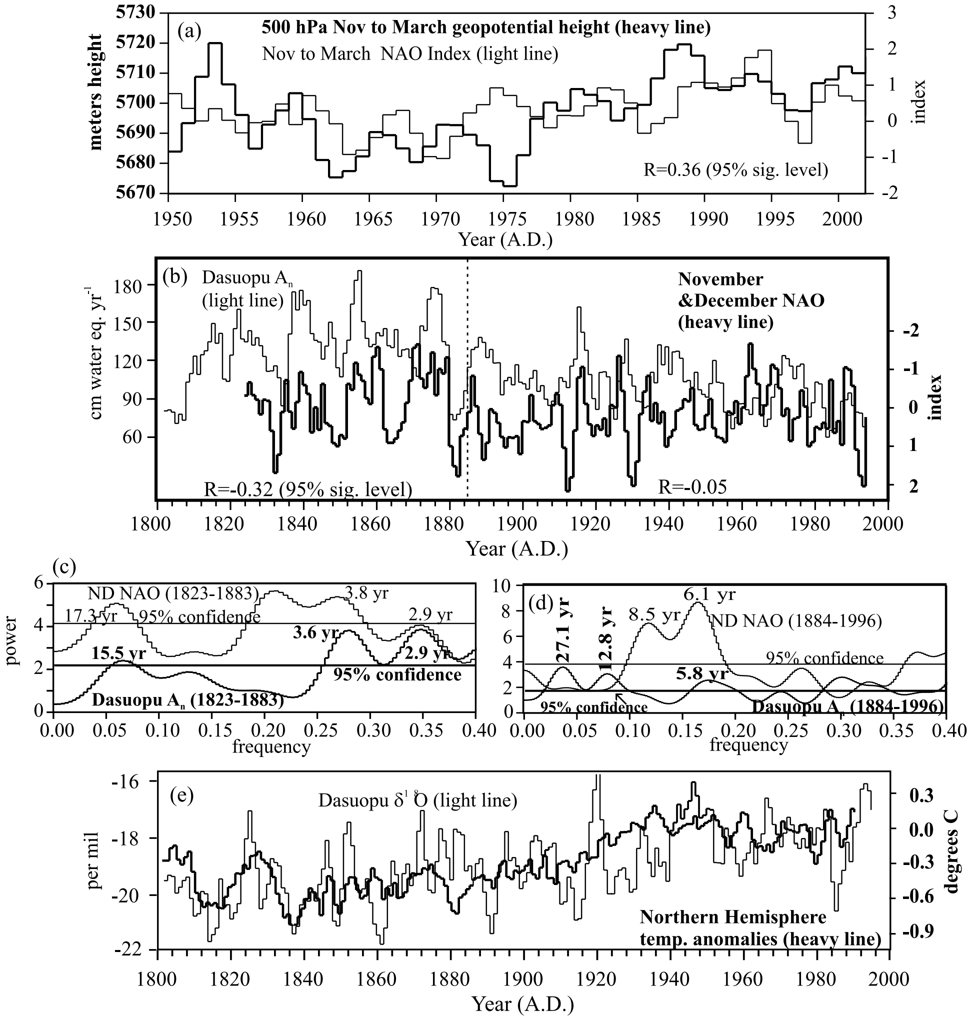

Figure 6. (a) Comparison between the North Atlantic Oscillation (NAO) Index (November to March, dark line) and the 500 hPa geopotential height (light line) from 1950 to 2003. (b) The Dasuopu A_n record (light line) is compared with the average November–December NAOI (dark line), and correlation coefficients before and after 1884 A.D. are shown. Power spectra of the NAO (light line) and Dasuopu A_n (heavy line) are shown for the period (c) from 1823 to 1883 and (d) from 1884 to 1996. The bandwidth used in the Blackman-Tukey analysis was 0.071 for Dasuopu (1810 to 1883), 0.83 for NAO (1823 to 1883), and 0.043 for the 1884 to 1996 series. The 95% confidence levels and significant periods (frequencies) are depicted for each series. (e) Comparison of Dasuopu $\delta^{18}O$ time series with the Northern Hemisphere temperature anomalies of *Jones et al.* [1998]. All the time series in Figures 6a, 6b, and 6e were smoothed with three-year running means.

15.5 to 17.3 years, 3.6 to 3.8 years and 2.9 years (Figure 6c). After the drop in accumulation and the increase in the NAOI at the end of the LIA, the power spectra diverge (Figure 6d). The Dasuopu A_n, which at this point reflects summer monsoon precipitation, maintains its quasi-decadal frequency, but also acquires a 27-year cycle that may be suggestive of Pacific Basin processes such as the Pacific Decadal Oscillation (PDO). The only real agreement during the post-LIA period is at 5.8 to 6.1 years, which suggests an ENSO signal. This lends support to the previous discussion that North Atlantic processes were influential over the early winter precipitation during the LIA, and later were replaced by tropical Pacific influences as the recent warming got underway and the monsoon circulation asserted more control over the annual Dasuopu accumulation.

[24] It is paradoxical that meteorological conditions in the late autumn and early winter would dominate the Himalayan snowfall in the later part of the LIA, since today the majority of the winter storms appear from January to March. However, the climatic conditions in the central Himalayas, and indeed over the Northern Hemisphere, during the nineteenth century were much colder than they have been since (Figure 6e). Limited documentation exists concerning the later onset of winter in Asia since the end of the LIA [*Shimaraev et al.*, 2002].

[25] It is unlikely that even large and persistent changes in the North Atlantic circulation during the late autumn-early winter would have been the sole cause of such large increases in winter snow in the central Himalayas. Orographically induced local storms, driven by cold

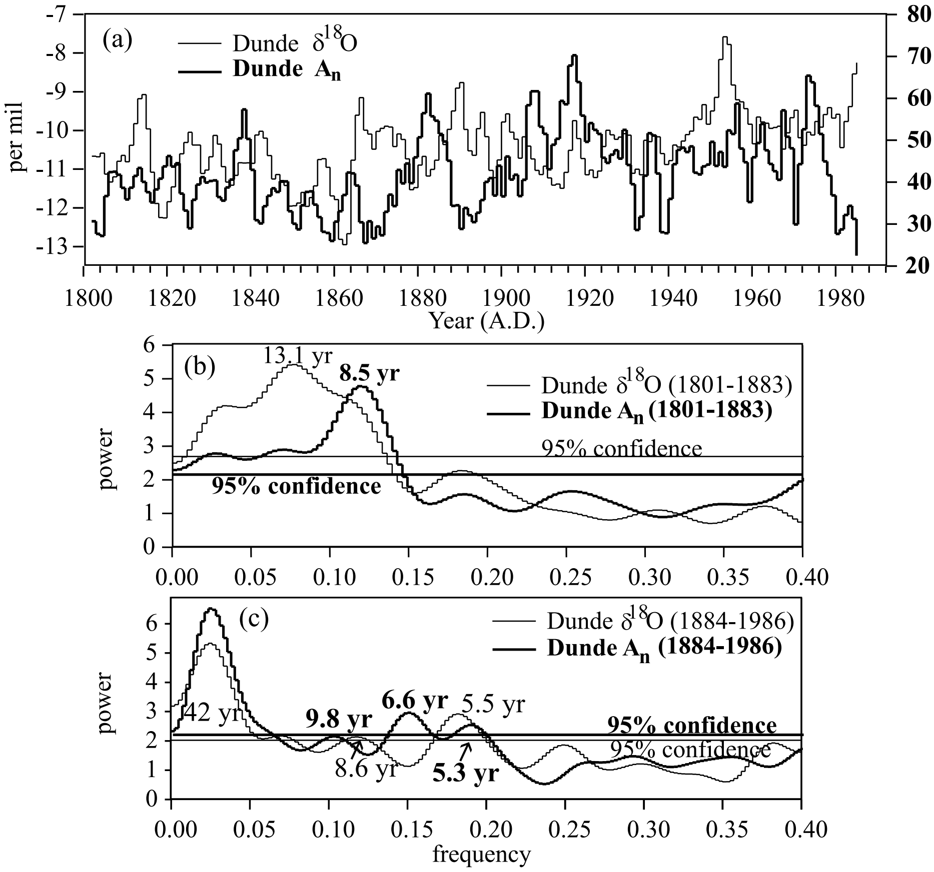

Figure 7. (a) Comparison of the Dunde ice core A_n (heavy line) and $\delta^{18}O$ (light line) records, both smoothed with three-year running means. Power spectra for the Dunde parameters are shown for the period (b) from 1801 to 1883 and (c) from 1884 to 1986. Significant frequency bands (over the 95% confidence levels) are noted. The bandwidth in the analysis was 0.047.

temperatures and the unusual upper level pressure patterns, may have been more frequent and intense. Also, more winter snow may have persisted into the summer half of the year than it has since the post-LIA warming began. Regardless of the extent of North Atlantic involvement in the climate of the Himalayas (and the rest of the monsoon regime), evidence does exist that a unique set of atmospheric conditions existed during most of the nineteenth century that occurred along with unusually heavy winter snow accumulation in this region.

5. Controls on Snow Accumulation on the Dunde Ice Cap in the NE Tibetan Plateau

[26] Currently an important, but not the only, precipitation source for the northeast Plateau in the summer is recycled moisture from the southeast Asian monsoon through the extensive Tibetan lake system. Because of the east-west orientation of the mountain ranges, monsoon precipitation can penetrate into eastern Tibet [*Sobel et al.*, 2001], but the steeply declining gradient of the moisture over distance insures that the amount that reaches the Dunde ice cap is small as reflected by the average annual accumulation rate (44 cm w.e.), some of which is deposited in the winter. The Dunde and Dasuopu A_n profiles are nearly mirror images of each other [*Davis and Thompson*, 2004]. When Dasuopu A_n was high during the nineteenth century, Dunde was low and their trends reversed through the recent warming. Does this suggest that the climatic processes affecting these two glaciers are linked?

5.1. Dunde A_n and Asian Land Temperatures Since the Little Ice Age

[27] Since a significant portion of Dunde's annual precipitation is monsoon-derived, its net accumulation record should show connections with the end members of the gradient between Asian land temperatures and tropical ocean SSTs. For the land temperature side of this relationship the Dunde $\delta^{18}O$ time series is useful, especially since the A_n and $\delta^{18}O$ chronologies are derived from the same ice core and therefore dating uncertainties that occur between independently constructed chronologies can be avoided. In Figure 7a, the records are smoothed with three-year running means to mask the high frequency noise. Simple visual inspection of the curves indicate that the relationships during the LIA are opposite (low/high temperatures, high/low accumulation) and continue thus until the early twentieth century. From 1800 to 1860 there is a strong and regular oscillation in both the parameters, although the period of the isotope oscillation is greater than that of the accumulation (13 versus 8.5 years, respectively, Figure 7b). Around 1900 to 1910, 20 to 30 years after the Northern

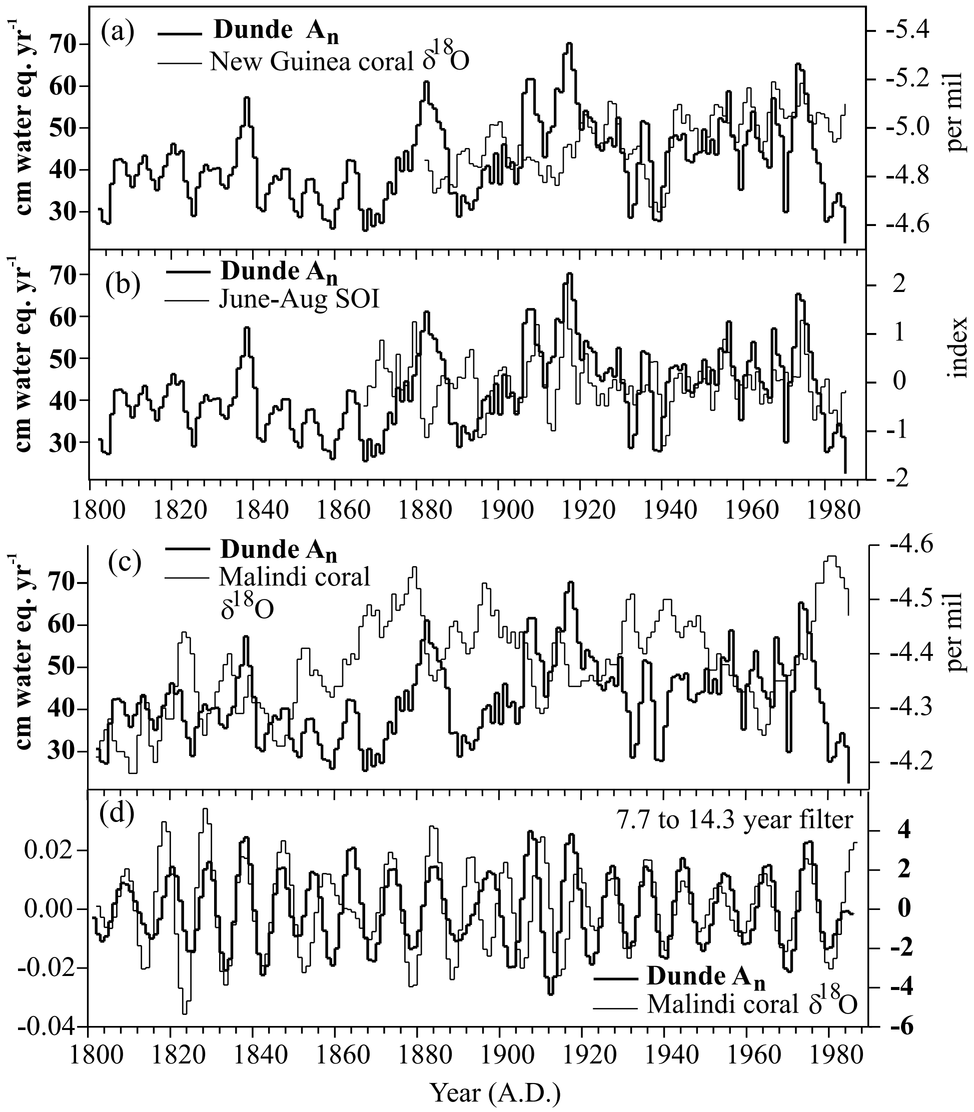

Figure 8. Comparison of Dunde A_n record (heavy line) with (a) a coral $\delta^{18}O$ record from Papua New Guinea, (b) the Southern Oscillation Index (SOI) from June to August, and (c) the Malindi coral $\delta^{18}O$ record from the western Indian Ocean. All time series are smoothed with three-year running means. (d) Comparison of filtered Malindi $\delta^{18}O$ and Dunde A_n series, both smoothed with a 7.7- to 14.3-year Gaussian band-pass-filter (bandwidth equal to 0.03).

Hemisphere warming began and the Himalayan accumulation abruptly decreased, high A_n was more often than not contemporaneous with high Asian surface temperatures, and the spectral frequencies of the two records show some agreement (Figure 7b). Both have a strong 42-year cycle, a quasi-decadal signal, and an ENSO-type signal at 5.3 to 5.5 years. The changing relationships between the isotope-derived temperatures and the reconstructed monsoon precipitation in this region suggest that the climate in the northeastern Tibetan Plateau underwent a mode shift between the LIA and the recent warming.

5.2. Dunde A_n and Tropical Sea Surface Temperature Records From Corals

[28] There are many high-resolution records of SST variations available from the equatorial Indian and Pacific Oceans, and those from living corals are among the most useful. One drawback for coral records, like those from ice cores, is the limited spatial coverage which is governed by site suitability and logistical considerations. A region of interest where records are available is Papua New Guinea, which is located close to the interaction between the southeast Asian monsoon and Pacific Basin processes such as El Niño-Southern Oscillation (ENSO). Unfortunately the longest continuous published record from this site (DT91-7) only extends back to 1880 [*Tudhope et al.*, 2001], and thus does not cover the LIA.

[29] If the New Guinea isotopic data reflect SST at the western end of the ENSO dipole and if the source for the Dunde summer precipitation is the Asian monsoon, then the coral and ice core records should show a relationship that reflects the connection between these ocean-atmosphere linked processes. From 1880 (the beginning of the coral record) to 1920, the correlation is insignificant (Figure 8a).

However, within the period since the end of the LIA that covers the recent warming, the correlations show some significance (depleted $\delta^{18}O$ with high A_n) from 1920 to 1986 (-0.39 at 95% significance level). Cooler than normal SST's at the western end of the ENSO dipole are characteristic of El Niño conditions, during which Asian monsoons have shown decreased strength. In order to confirm the degree to which the ice core record reflects ENSO variability, a comparison with the June through August (JJA) Southern Oscillation Index (SOI) is illustrated in Figure 8b. Since the middle 1890s, the correlation between Dunde A_n and SOI has been significantly positive (+0.50, 99% significance level) and all three records show a strong 6.6-year signal which is visually obvious, but the relationship was reversed before ~1895 (-0.42, 95% significance level). Unfortunately as with the coral, the SOI record does not extend back early enough to observe the Pacific Basin conditions during the coldest part of the LIA.

[30] One of the few corals that provide a continuous isotopic record since the beginning of the nineteenth century is from the Malindi Marine Park in the western equatorial Indian Ocean. Visual inspection of the comparison between Malindi $\delta^{18}O$ and Dunde A_n (Figure 8c) shows a strong positive correlation (1881 to 1986, $R = 0.49$, significance at 99%), i.e., enriched/depleted isotopes (cold/warm SST) are coincident with high/low accumulation. Before 1880, the correlation is nonexistent. *Cole et al.* [2000] noticed a strong decadal signal in this coral from which they concluded that climatic variability in the western Indian Ocean is heavily influenced by the tropical Pacific. Similarly, a 9.8-year signal is observed in the Dunde accumulation record from 1881 to 1986 (Figure 7c). In order to isolate these decadal signals, both time series were smoothed with a 7.7 to 14.3-year band-pass-filter (Figure 8d). Alignment between the coral and the ice core is obvious from 1920 to the early 1980s, which to be expected if the Dunde accumulation is reflecting Pacific oscillations during the twentieth century warming. In fact, in Figure 8a the coral record from New Guinea and Dunde A_n begin to show greater correlation around 1920. Therefore, during the recent warming trend, after the seemingly pivotal period in the late 1880s, the Dunde accumulation time series seems to reflect the interlinked Pacific-Indian Ocean processes that are documented in the coral records [*Cole et al.*, 2000], along with the Indian Ocean Dipole [*Saji et al.*, 1999], which is thought to influence the Asian monsoon [*Charles et al.*, 2003].

[31] Figure 8d also shows two short periods of alignment in the nineteenth century, from the mid-1820s to 1850, and from 1870 to 1890. It may only be coincidental, but during these two intervals the Dunde $\delta^{18}O$ are enriched (Figure 7a), which may indicate some degree of warming in the central Asia. Accumulation actually shows sharp increases in the middle of each of these intervals, but whether that is due to increased precipitation from the south is impossible to determine from these records.

6. Discussion

[32] Observational and model studies have presented evidence for a cause and effect mechanism between the extent and/or depth of late winter-early spring snow cover over various parts of Eurasia and the strength of the subsequent summer monsoon. Large amounts of energy are required to melt a larger spring snow pack, inhibiting the sensible heating of the high Asian region and the development of the thermal low pressure that is important for establishing the summer monsoon circulation. The Himalayas are thought by some to be an especially critical region in this mechanism [*Dong and Valdes*, 1998; *Kripalani et al.*, 2003].

[33] Linkages between the North Atlantic and winter storms in the Himalayas during the Little Ice Age prior to the end of the nineteenth century were discussed above. The persistence of the snow late into the year over several decades in the southern Tibetan Plateau may have been a causal mechanism for a weaker monsoon circulation. The extended all-India summer rainfall (AISR) record from 1813 (not shown) indicates that the middle of the nineteenth century was overall the driest in India [*Sontakke and Singh*, 1996] at the same time that the Dasuopu accumulation was at its highest. In fact, the Dasuopu data and the AISR record show opposite trends up until the middle 1880s, then switch into a positive relationship until the early 1960s. This supports the findings of *Robock et al.* [2003], who noted a changing correlation between the all-India rainfall record (as representing the strength of the Asian summer monsoon) and Eurasian snow cover (using the NAO as a proxy) such that the two were only significantly correlated in the 1880s and the late twentieth century, but these comparisons only extend back to 1871. During these two periods, lower NAO indices were coincident with low all-India rainfall (AIR), and it was also during these periods that interannual variability in the winter Eurasian circulation and the summer monsoon was highest. This presented a challenge to *Blanford* [1884], which promoted snow cover over the Himalayas as a consistent predictor of subsequent summer Indian monsoon strength. However, in the case of the Dasuopu record, it is proposed here that winter precipitation in the Himalayas was much higher up until ~1880 and added to the summer snowfall to result in a 30% increase in annual accumulation, and that cold temperatures allowed this deep snow to persist into the warm season, thus inhibiting the development of the subsequent summer monsoon. After the LIA ended, the winter snow totals in the central Himalayas abruptly dropped, the land and ocean temperatures began to increase while the monsoon strengthened, and the summer precipitation became more significant for Dasuopu A_n variability.

[34] Snow accumulation was low in the northeast Tibetan Plateau during the LIA, which may have been a result of weaker Asian monsoon circulation during this period of cold Asian land temperatures. The increase in Dunde A_n along with lower ice core isotope-inferred temperatures over decadal periods suggests that the climate here may have been more influenced by westerly low-pressure systems in the summer and especially in the winter. This continental-derived precipitation may have dominated as the monsoon circulation, which is relatively weak in this region even under climates that favor it, was restricted to the south. Today the northeastern Tibetan Plateau is on the extreme edge of the summer monsoon regime, so any weakening of the monsoon might place this area outside its influence.

[35] After 1880, the winter storms decreased in the central Himalayas as the effects of the North Atlantic atmospheric processes diminished and the regional and hemispheric climate ameliorated, facilitating spring melting of the snow cover in the lower elevations. The precipitation on the Dasuopu ice cap more resembled the summer monsoon rainfall record to the south. Meanwhile, to the north the Dunde A_n fluctuated greatly, increasing over three-fold and reaching a maximum at 1884, then falling back in six years to LIA levels, then climbing again to a record high by 1920. The west Pacific SSTs and the summertime Southern Oscillation settled into their monsoon relationships with the ice core record (high SST, positive SOI, high Dunde A_n) in the mid-1890s, more than a decade after the decrease in Himalayan winter snow. However, the Asian land temperatures did not achieve a positive relationship with monsoon accumulation until ~1910, which they have maintained until the 1970s. This suggests that after the end of the LIA as the recent warming got underway, the ocean processes may have been more influential in expanding the monsoon influence towards the north and west. In fact, even during the twentieth century the correlations of the ice core accumulation with the Pacific Basin atmospheric and oceanic processes are more significant than with the land temperatures on decadal and shorter timescales, whereas land temperature variations are more influential over multi-decadal and centennial periods.

[36] The appearance of post-LIA similarities between the Indian Ocean coral-inferred SSTs and the ice core-inferred south Asian monsoon strength demonstrates that Indian Ocean processes also influence the monsoon regime. The decadal signals from western Indian Ocean coral records reflect tropical ocean, particularly Pacific Basin, processes [*Charles et al.*, 2003]. Thus the south Asian monsoon and the Indian Ocean are linked through the tropical Pacific and ENSO.

7. Conclusions

[37] Changes in precipitation and temperature on the Tibetan Plateau which affect the extent and thickness of snow and ice have direct and indirect implications for water resources and agriculture in western China, Pakistan, India, and Southeast Asia, especially since major rivers such as the Yangtze, the Indus, the Ganges, and the Brahmaputra originate in Tibet and the Himalayas. The comparisons of temperature proxy records from the land and the oceans with ice core accumulation data are useful for reconstructing variations in the relative influences of these end members of the thermal gradient, which is the active monsoon's basic driver. Analysis of climate histories from ice core and coral proxy records of temperature and monsoon strength suggests that over the last century monsoon derived precipitation in the northeast Tibetan Plateau was associated more closely with tropical SST than with temperature changes over Asia on decadal and shorter timescales. However, during the Little Ice Age cold period in the nineteenth century North Atlantic and Eurasian atmospheric circulation and Asian land temperatures may have exerted more influence than did tropical processes. The Dunde and Dasuopu glaciers, located on the edges of the Tibetan Plateau, may be situated in critical regions where tropical and high latitude climate regimes compete as the Northern Hemisphere shifts in and out of neoglacial (and glacial) modes. This may be important to consider while reconstructing and interpreting the paleoclimate histories from longer ice core and other proxy records from the Plateau.

[38] Up until the middle 1970s, ENSO seemed to have an effect on monsoon strength, but since then the accelerating climatic warming may have resulted in land temperatures playing a more important role in the thermal gradient that drives the monsoon circulation [*Ashrit et al.*, 2001]. Unfortunately, this recent relationship cannot be reconstructed from Tibetan Plateau ice core records, since tropical and temperate glaciers in this region have been experiencing rapid and accelerating retreat, thus disrupting their recent climatic records.

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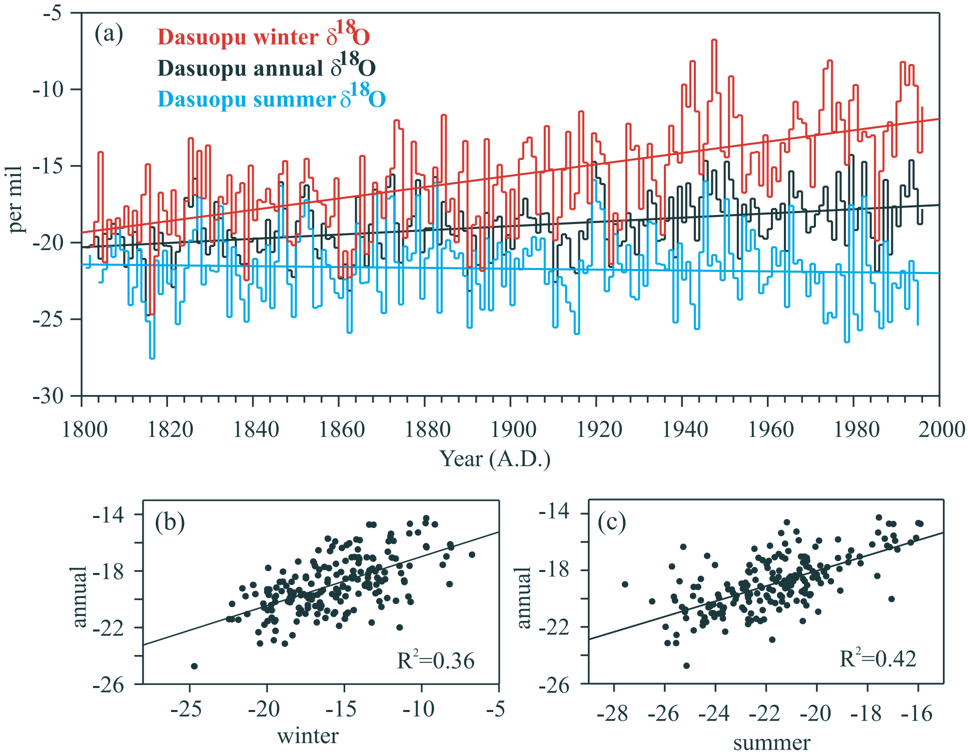

Figure 3. (a) Time series (annual averages) for Dasuopu $\delta^{18}O$, together with the winter seasonal values and summer seasonal values. The regressions between the (b) annual and winter and (c) summer and annual $\delta^{18}O$ time series are at the 99% significance level, N = 195). None of the data are smoothed.