27 Ice core evidence from Peru and China

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27.1 Introduction

Ice sheets and ice caps are recognized as being the best of only a few sources of atmospheric history from which past climatic and environmental conditions may be extracted. Much of the climatic activity of significance to humanity may not be strongly expressed (or may not affect) the polar ice caps. Ice core records can be recovered from polar ice sheets, as well as from a select few high altitude, low- and mid-latitude ice caps. In addition, the high accumulation on these ice caps and ice sheets makes it possible to recover high temporal resolution records of particulates, chemical constituents and oxygen isotope ratios ($\delta^{18}O$) for the last 1000 years. The aerosol records (soluble and insoluble) records serve as indicators of drought, wind strength, volcanic activity, and net accumulation, while $\delta^{18}O$ can, in some cases, serve as a proxy for temperature. Dust plays a deterministic role in the atmospheric radiation balance. Thus, the variety of chemical and physical data extracted from ice caps and ice sheets provide a multi-faceted record of both the climatic and environmental history of the earth and may allow assessment of the relative importance of potential forcing functions such as volcanic activity, greenhouse gas concentrations, atmospheric dust, and solar variability.

One of the important needs in addressing global change issues which the ice core records can provide are long time series, that is, a frame of reference against which present and future changes can be compared. The longer-term perspective available from polar cores is well documented. Recent evidence (Thompson *et al.* 1989a) indicates that longer-term glacial-interglacial records from non-polar regions can be obtained from the Qinghai-Tibetan Plateau. Such records provide a more global perspective of climate which is needed to understand fully the earth's climate system, and specifically to determine how well polar ice cores reflect climate variability in the subtropics. This chapter presents ice core evidence of climatic and environmental variability since A.D. 1500 with emphasis on those records from the tropical Quelccaya ice cap, Peru, and the subtropical Dunde ice cap, China (Figure 27.1).

The Tibetan Plateau and Bolivian-Peruvian altiplano have similar mean elevations of approximately 4000 meters. The regional weather patterns are driven by the sensible heat flux over these plateaus and the subsequent latent heat release during precipitation. Gutman and Schwerdtfeger (1965) studied the role of latent and sensible heat for the development of the Bolivian High and concluded that similar conditions exist there as in the highlands of Tibet (Flohn 1965). In both cases the upper troposphere anticyclone is a quasi-stationary system, persisting throughout the warm season. A typical weather pattern over the plateau during the wet season is a clear morning during which the plateau is heated, followed by an afternoon thunderstorm. Over 80% of the annual snowfall on the Quelccaya ice cap, Peru and the Dunde ice cap, China falls during the wet season (November-April) and (April-August) respectively. Since the Tibetan High is an integral part of the Asiatic monsoon circulation during the northern summer, it is instructive to compare not only the Bolivian and

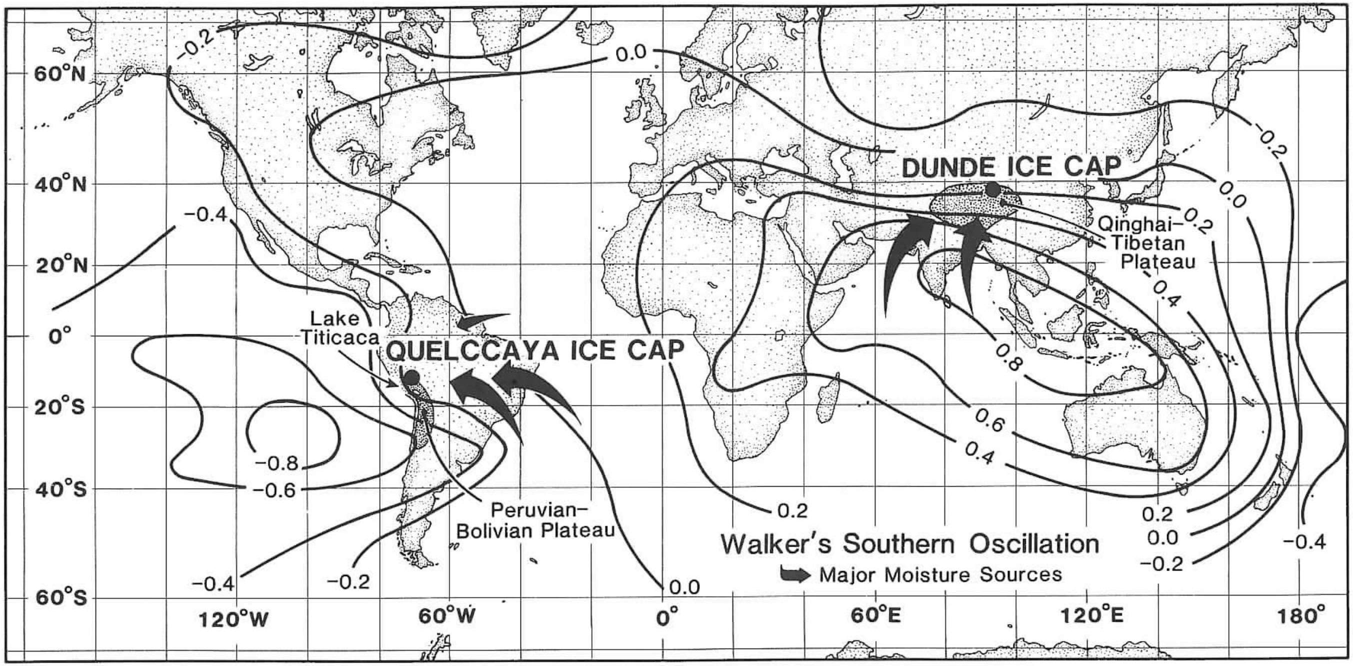

Figure 27.1 Location map showing the position of the Quelccaya ice cap, south of the equator on the Bolivian-Peruvian altiplano and the Dunde ice cap, north of the equator on the Tibetan Plateau. Arrows indicate the major moisture sources for both locations. The Walker circulation pressure correlations illustrate how teleconnections could link these two widely separated ice caps.

Tibetan Highs, but also other major features of both the upper and lower level circulation over South America and the Asiatic summer monsoon circulation.

Virji's (1979) comparison between the features of the Asiatic summer circulation and that over South America during the southern summer, suggests that the systems are similar in a number of important respects, and that the character of the summertime circulations over tropical and subtropical South America is monsoonal. Virji notes that the circulation system over South America is relatively more compact, confined largely to the continental zone. In contrast, the Asiatic monsoon system embraces a much larger area including the Indian Ocean and the adjacent continental region of Africa. Despite obvious differences in areal extent and intensity of the various features, the two circulation systems resemble each other as follows:

(a) The quasi-stationary Bolivian and Tibetan Highs have maximum amplitudes near the 300mb level (Dean 1971; Krishnamurti and Bhalme 1976). Both are localized regions of intense Hadley-type meridional, as well as seasonal, east-west circulations.
(b) The well-marked upper level easterly flow around 10°S latitude over South America occasionally produces high speeds, analogous to the tropical easterly jet of the Asiatic summer monsoon system around 10°N latitude.
(c) The continental heat low over Gran Chaco and the Pampean Sierras between 15° and 30°S and 60° and 68°W just east of the Andes is a relatively weaker feature compared with the monsoon trough over northern India.

27.2 Quelccaya ice cap

Research programs were conducted on the tropical Quelccaya ice cap in the Peruvian Andes (13°56′S; 70°50′W) between 1974 and 1984. This relatively large ice cap is characterized as follows: summit elevation, 5670m; total area 55 km^2; mean annual temperature, -3°C; maximum summit ice thickness, 164m; flat bedrock topography; and net annual accumulation, 1.15m of water per year. The annual cycle in precipitation is characterized by 80-90% of the snow falling from November to April (Thompson *et al.* 1984a). This produces the distinct seasonality in precipitation which is subsequently preserved in the ice stratigraphy.

Because the site is so remote and too high for use of a conventional drill system, a newly designed, portable, lightweight, solar-powered drill was used in 1983 to recover two ice cores (163.6m to bedrock and 154.8m) without contaminating the pristine environment or the core samples. This was the first major ice core drilling project using solar power.

The annual layers for the past 1500 years were counted using a combination of visible seasonal dust layers and the seasonal variations of $\delta^{18}O$, microparticles and liquid conductivity, thus allowing a very precise time scale to be established. Total particle (diameter >0.63μm) and large particle (diameter >1.59μm) concentrations, liquid conductivity, $\delta^{18}O$, net accumulation and pollen have been measured (Thompson *et al.* 1986; 1988a).

Figure 27.2 illustrates profiles of total particle concentrations, liquid conductivity and $\delta^{18}O$ measurements for the Quelccaya ice cap, plotted with depth. Each of the parameters show significant variations with depth. In order to document changes in conductivity which may occur due to storage time in which water is bottled before analysis in the laboratory, 1000

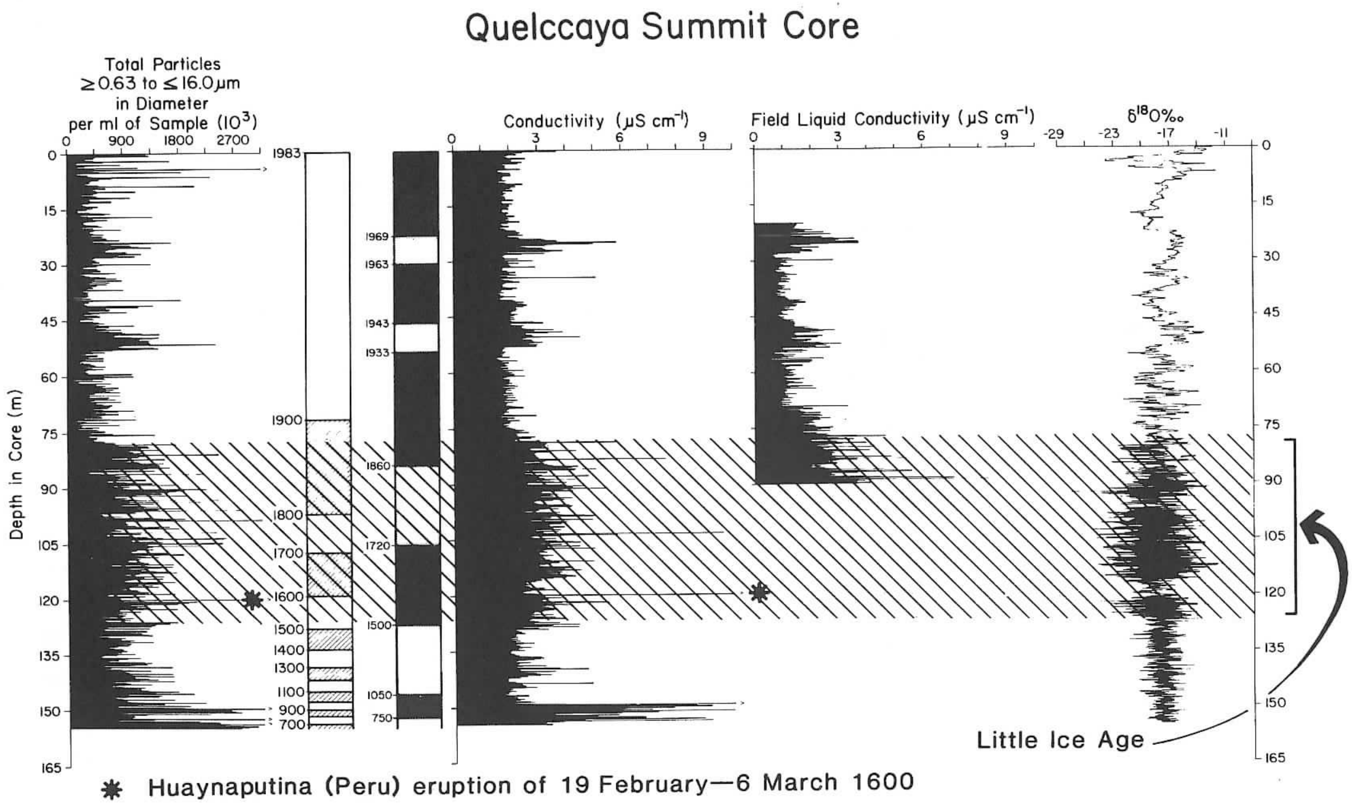

Figure 27.2 Overview of Quelccaya core 1 results; total microparticles, liquid conductivities measured in the field and oxygen isotopes plotted with depth in core. The second column is the stratigraphic time scale illustrating the compression of time with depth. The third column indicates the wet and dry periods as reported in Thompson *et al.*, (1985). These data provide a continuous record of the environmental changes since ~A.D. 470. The hatched area indicates the Little Ice Age (LIA).

conductivity measurements were made in the field immediately after the ice core was drilled, and then remeasured several months later when the samples arrived at the laboratory. The results illustrated in Figure 27.2 demonstrate that although background levels increase due to dissolution of some particles and absorption of atmospheric CO_2 with time, the major features are reproducible. The second column in Figure 27.2 illustrates how the rate of change in age increases with depth due to compression and thinning of the annual layers as a result of ice flow. The third column illustrates wet periods (dark shading) and dry periods (light shading) as calculated from the measured annual layer thickness. These data illustrate that there is no constant relationship between dry periods and increased concentration of either soluble or insoluble dust.

27.3 Dunde ice cap

Research programs have been conducted on the subtropical Dunde ice cap (38°06′N; 96°25′E) from 1984 to present. This ice cap is characterized as follows: summit elevation, 5325m; total area, 60 km^2; mean annual temperatures, -7°C (as determined from measuring 10m temperatures and is considered a maximum); maximum summit ice thickness, 140m; flat bedrock topography; and net snow accumulation, 0.2m of water per year. The record of dust is very promising as the Asian continent is a major source area for continental dust (e.g., Gobi Desert) which contributes substantially to the North Pacific atmospheric dust burden today and in the past.

As with Quelccaya, over 80% of the snow falls in the wet season (monsoon season, May-August) producing the marked visible dust stratigraphy. In 1987 three ice cores, 136.6, 138.4 and 139.8m in length, were drilled to bedrock at the summit of the ice cap. Borehole temperatures indicate that even though the ice cap is located 38°N, it is a "polar" glacier with a bottom temperature of -4.7°C, well below the pressure melting point (Thompson *et al.* 1988b; Thompson *et al.* in press).

27.4 Annually resolvable ice core records

Fortunately, the high altitude regions in the low and mid latitudes, as well as the polar regions, contain ice caps and ice sheets which have continuously recorded climatic and environmental changes over the past several thousand years, often with annual resolution (Dansgaard 1954; Johnsen *et al.* 1970, 1972; Thompson 1977; Thompson *et al.* 1986; Mulvaney and Peel 1988; Hammer 1989). The records can be used to address problems which are of concern to a wide sector of the scientific community, governments, industries, and the general public. These include: (a) global scale climatic events such as the Little Ice Age (LIA) and El Niño/Southern Oscillation (ENSO) and monsoonal variability; (b) abrupt climatic change and evidence for changes in the amplitude of the annual cycle over the last 1000 years; (c) impact of past low and high frequency climatic changes on human activities, and (d) documentation of climate changes in the 20th century relative to long time-perspectives provided by ice core records.

The construction of a time scale for an ice core is generally based upon the integrated high resolution (8 to 12 samples per year) records of $\delta^{18}O$, microparticle concentrations, conductivity, and ionic concentrations which often exhibit a distinct seasonal variation in flux (Thompson *et al.* 1986; Hammer 1989). When possible, as is the case for many high elevation non-polar ice caps, visible dust layers allow rapid and accurate dating (Thompson *et al.* 1985). The measurement of total Beta radioactivity makes it possible to identify the major global time-stratigraphic horizons associated with known atmospheric thermo-nuclear tests (Picciotto and Wilgain 1963; Lambert *et al.* 1977, Thompson *et al.*, in press). Two or more ice cores should be drilled at the same site and similarly analyzed to provide an independent verification of the time scale and to ensure an uninterrupted physical and chemical stratigraphic record. For most of the ice core records presented here, at least two, and in some cases three, cores are available from each site. The utility of ice core analyses hinges upon accurate dating of cores which requires the use of multiple stratigraphic features exhibiting a seasonal variation and/or visible annual dust layers which are preserved within the ice. Figure 27.3

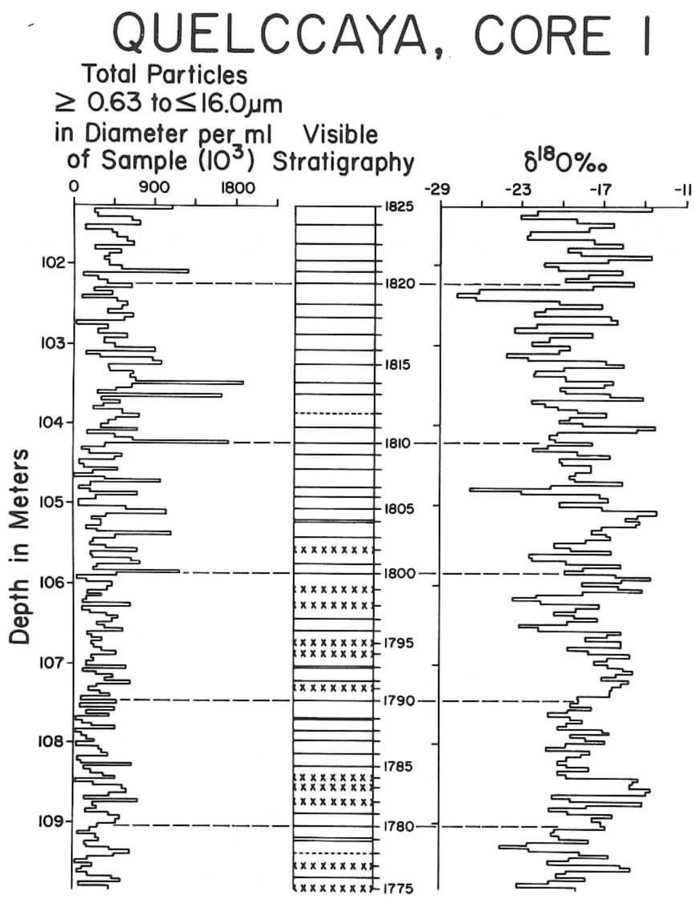

Figure 27.3 The 50-year period from A.D. 1775-1825 illustrates stratigraphic parameters used to date the Quelccaya cores. Annual signals are recorded in microparticle concentrations (diameters $\geq 0.63 \leq 16.0 \mu m$ per ml of sample) oxygen isotope ratios and visible stratigraphy. For stratigraphy, a single solid line represents a normal dry-season dust layer; a single dashed line and double dashed line represent very light and light dust layers, respectively. Series of X's symbolize diffuse dry season layers (Thompson *et al.*, 1986).

illustrates a 50-year period A.D. 1775-1825, from the Quelccaya Ice Cap. Distinct annual layers are preserved in particles, visible stratigraphy and $\delta^{18}O$ and permit very accurate dating of the ice cores (Thompson *et al.* 1986). On the Quelccaya ice cap the least negative oxygen isotopes, highest concentration of insoluble and soluble dust and the visible dust layers occur during the dry season – May through October. Figure 27.4 illustrates the detailed stratigraphy preserved in particle concentrations and $\delta^{18}O$ in snow pits from the summit of the Dunde ice cap, China. The average measured accumulation rate from 31 stakes in the summit strain network for the 1986-1987 accumulation year was 80cm of snow or 0.34m of water equivalent. On the Dunde Ice Cap high dust concentrations occur during the late fall to early spring, prior to the onset of the monsoon snowfall, while the least negative $\delta^{18}O$ values occur in August and September after the monsoon season. The low average $\delta^{18}O$ values and large seasonal variations in $\delta^{18}O$ in snow accumulating on the Quelccaya ice cap can be quantitatively explained as a result of atmospheric processes between the oceanic source of water vapor and Quelccaya that determine the isotopic composition of snow deposited on the ice cap, and local conditions determining how the depositional isotope signal is modified during firnification (Grootes *et al.* 1989). However, this explanation for the observed $\delta^{18}O$ variations is probably not unique as it does not consider the very large vertical height of convective thunderstorms in the Amazon Basin which are often two to three times higher than the Andes. Thus, the observed $\delta^{18}O$ in precipitation is the result of condensation at many different temperatures throughout the vertical column. The comparison of the measured accumulation with the dust and oxygen isotope records confirms the annual cycle in these parameters.

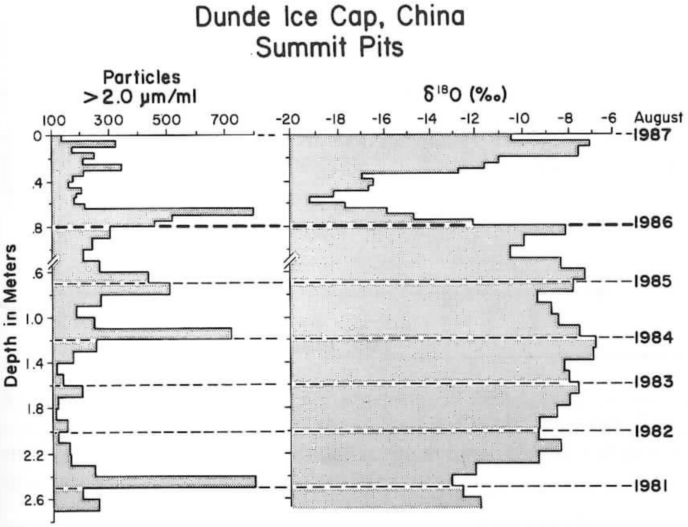

Figure 27.4 The stratigraphic parameters used to date the Dunde cores are illustrated for two snow pits, sampled in 1986 and 1987. The 1986-87 accumulation was determined by measuring accumulation stakes. Annual signals are recorded in microparticle concentrations (diameters $\geq$ 2.0 and $\leq$ 60$\times\mu$m per ml of sample) and oxygen isotope ratios. The // symbol indicates where the 1987 and 1986 snow pit records are joined.

Table 27.1 presents the correlations of decadal averages for Quelccaya ice cap core 1 and summit core for accumulation, $\delta^{18}O$ and dust for the period A.D. 1570-1980. The $\delta^{18}O$ records are most strongly correlated and the dust records are least well correlated. Table 27.1 illustrates that the reproducibility of any given chemical constituent of the record is partially a function of the parameter being measured. Stable isotopes are more highly correlated in part because they undergo smoothing due to vapor and molecular diffusion during the firnification process. Thus, some of the "noise" or detail in the record is smoothed out and lost in the process. A comparison of the annual oxygen isotope data between core 1 and summit core for the period A.D. 1500-1983 yields an r value of 0.87, significant at the 99.9% level only slightly less than the decadal correlation of 0.95. Particles, on the other hand, represent physical entities within the ice and are not subject to diffusion. Moreover, the position of the drill site upwind or down wind of the summit could greatly influence the concentration of dust. Differences may also result from the fact that the measured annual layers at the core 1 site average 19cm over the past 1000 years versus 18cm at the summit core site. The difference in average layer thickness is a result of the different flow regimes and vertical strain rates at the two sites (Thompson *et al.* 1985). Thicker individual annual layers contain a greater volume of water equivalent for the same year which produces a lower particulate count per unit volume per sample.

Table 27.1 Statistical correlations of decadal averages for Core 1 and Summit Core ice core parameters for the period A.D. 1570-1980.

	r	r^2
Accumulation	.89*	79%
Oxygen isotope ratio	.95*	90%
Microparticle concentration	.56*	31%

*significant at 99.9% level

In areas where it can be demonstrated that little or no mass loss occurs due to melting or removal by wind, ice cores provide the very best record of the past variations in precipitation available. The annual accumulation rate can be determined by measuring layer thickness changes based on annually varying ice core parameters such as dust, $\delta^{18}O$ and ionic concentrations. Records covering 1500 years have been produced from the Greenland ice sheet (Reeh *et al.* 1978) and from the tropical Quelccaya ice cores (Thompson *et al.* 1985). The annual records of annual layer thickness from Quelccaya are compared in Figures 27.5-27.7

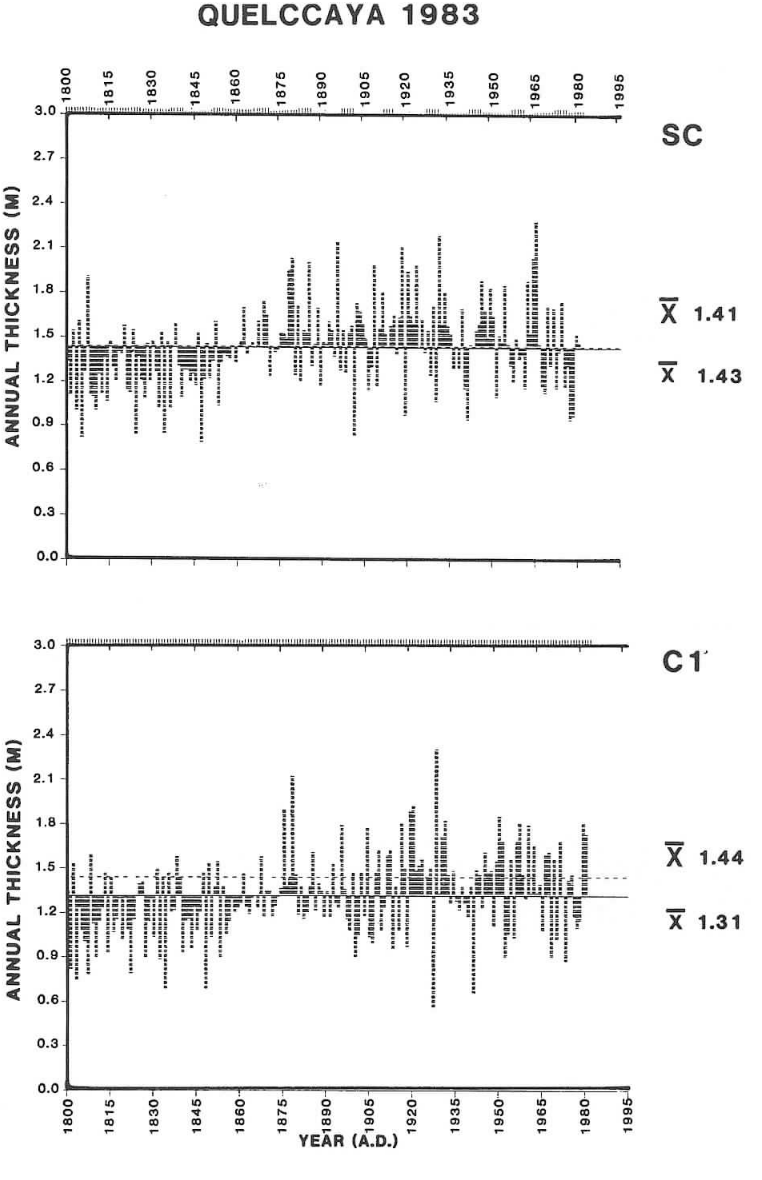

Figure 27.5 Annual layer thicknesses were determined in the summit core (SC) and core 1 (C-1) for the period A.D. 1800 to 1980. Large $\bar{X}$ is the average for entire core, small $\bar{x}$ is the average for the 180 year period illustrated.

which illustrate substantial annual variability, due in part to the effects of drifting and to drill site position on the ice cap.

On the Quelccaya ice cap the annual net accumulation record of A.D. 1915-1984 compares well with annual changes in Lake Titicaca water levels and with annual precipitation at El Alto (La Paz, Bolivia) suggesting that the 1500-year net accumulation record from Quelccaya may serve as a proxy for water level changes in Lake Titicaca (Thompson *et al.* 1988a). The

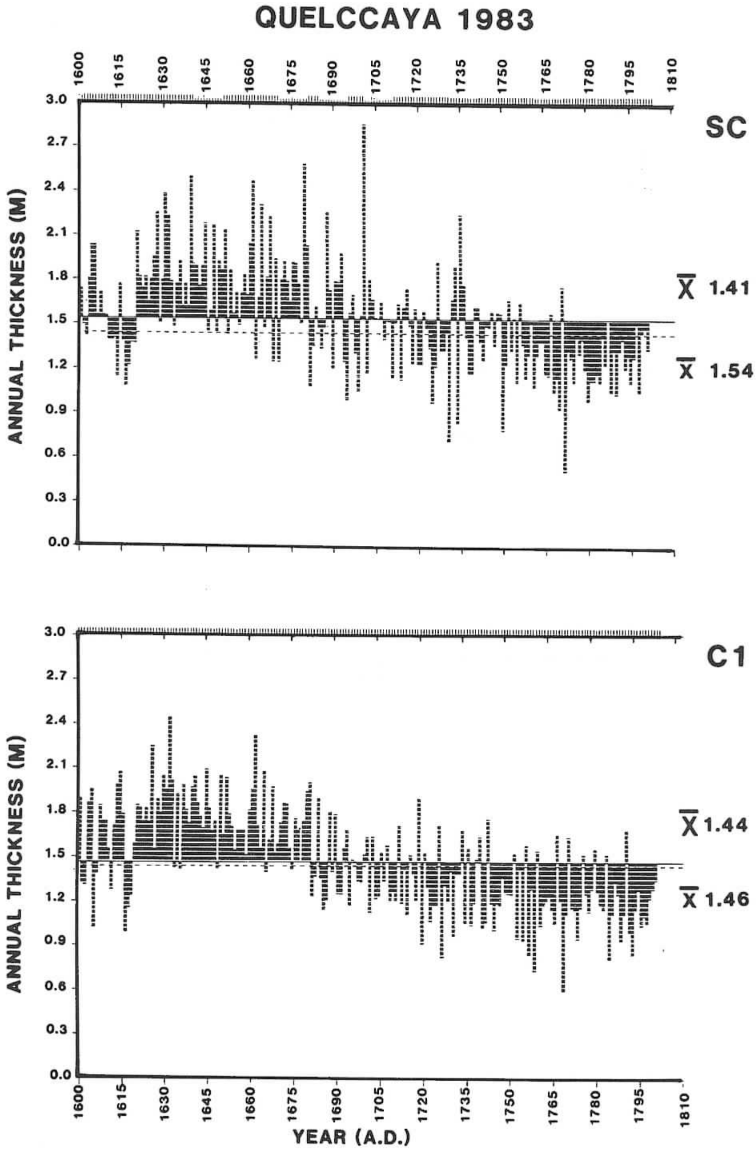

Figure 27.6 Identical to Figure 27.5, but representing the period from A.D. 1600 to 1800, except the small x̄ is the average for the 200-year period illustrated.

ice-core record suggests that climatic variability, reflected in lake level changes, has strongly influenced fluctuations in agricultural activity in southern Peru (Thompson *et al.* 1988a). As historic and prehistoric highland civilizations in Peru, Ecuador and Bolivia were largely agrarian based, and since the high plateau areas (being at the upper limits of agriculture) are climatically sensitive (Cardich 1985) it is likely that climate played an important role in the survival and economic well being of these cultures (Thompson and Mosley-Thompson 1989). The accuracy of the precipitation reconstructions, particularly in the deeper core sections,

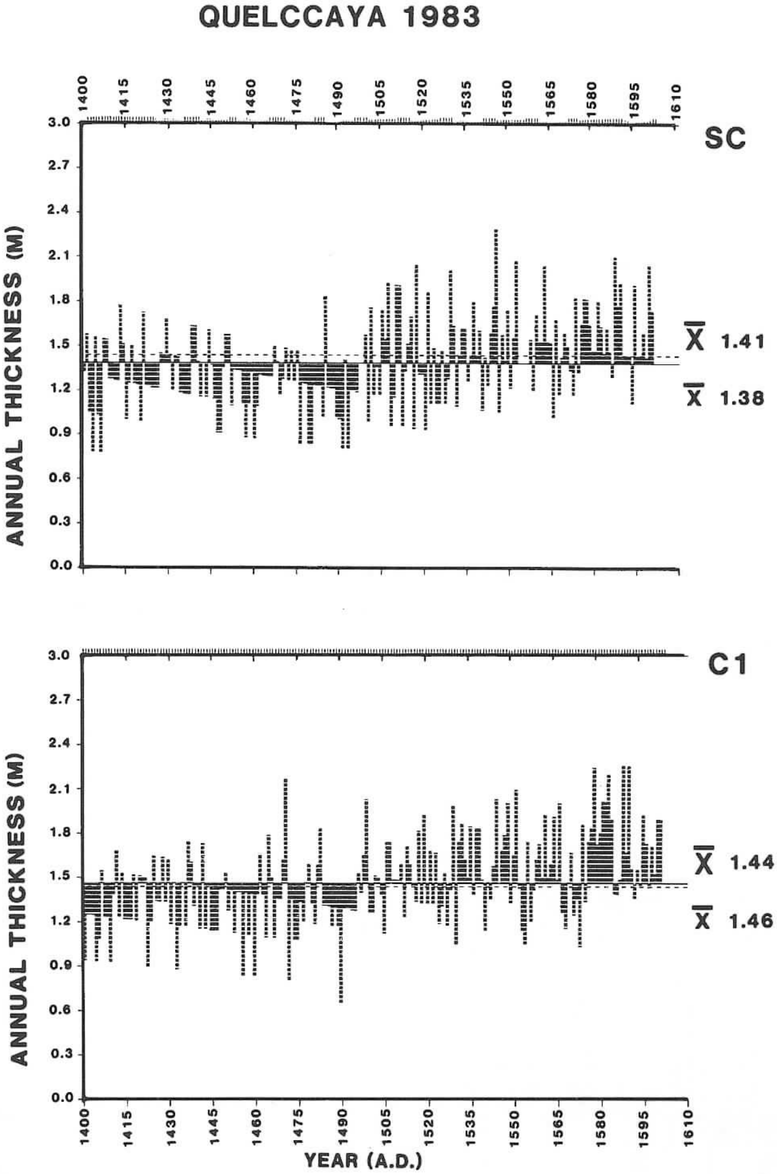

Figure 27.7 Identical to Figure 27.5, but representing the period from A.D. 1400 to 1600, except the small $\bar{x}$ is the average for the 200-year period illustrated.

depend on the application of appropriate constraints to the ice flow model (Paterson and Waddington 1984). Comparisons between accumulation (Figures 27.5-7) $\delta^{18}O$ (Figures 27.8-10) and particulates (Figures 27.11-13) illustrate that caution must be exercised in the interpretation of climatic and environmental history from a single ice core record. Figures 27.5, 27.6 and 27.7 illustrate that in both Quelccaya ice cores (1) A.D. 1400 to 1500 was a period of reduced annual layer thickness, (2) starting ~A.D. 1500 and lasting until

~A.D. 1720 was a period of enhanced annual layer thickness, (3) A.D. 1720 marks the beginning of a long period of reduced annual layer thickness which lasted until about A.D. 1860 and (4) the period of transition between this period of reduced annual layer thickness which characterized the 18th and 19th century and the slightly above average annual layer thickness of the twentieth century is marked by a transitional period A.D. 1860 to 1875 characterized by very little interannual variability in layer thickness. Figures 27.8, 27.9 and

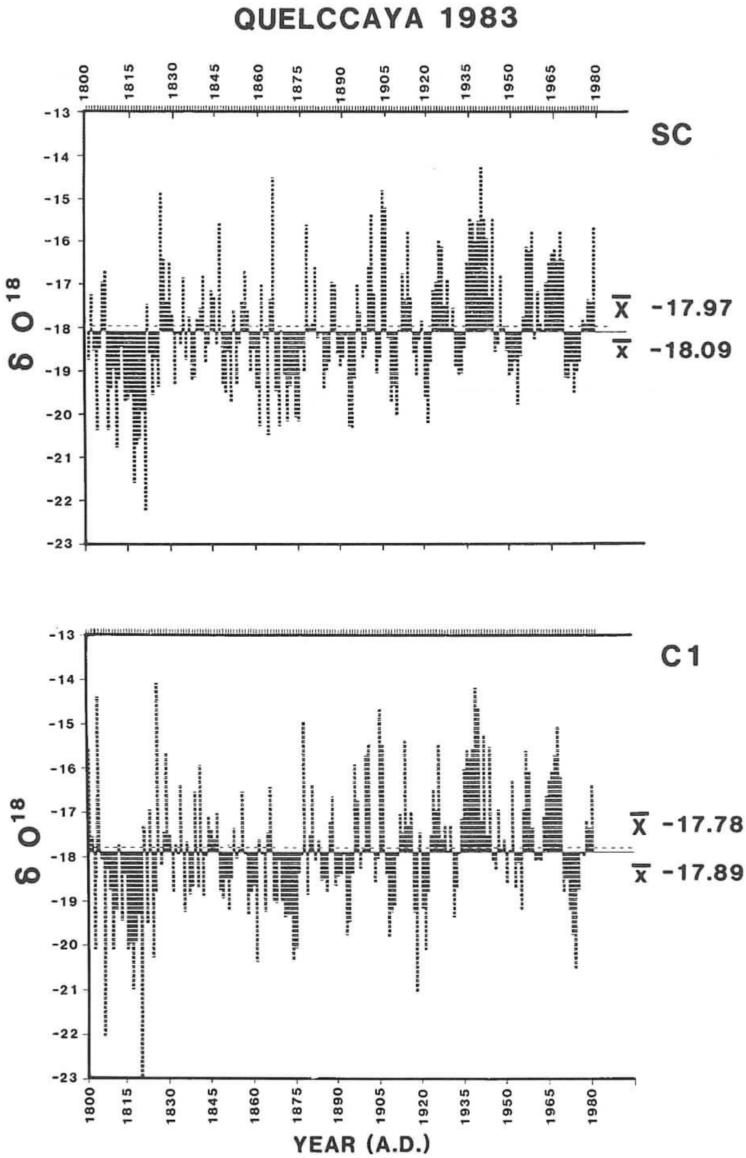

Figure 27.8 Annual $\delta^{18}O$ values determined in the summit core (SC) and core 1 (C-1) for the period from A.D. 1800 to 1980. Large $\bar{X}$ is the average for the entire core, small $\bar{x}$ is the average for the 180 year period illustrated.

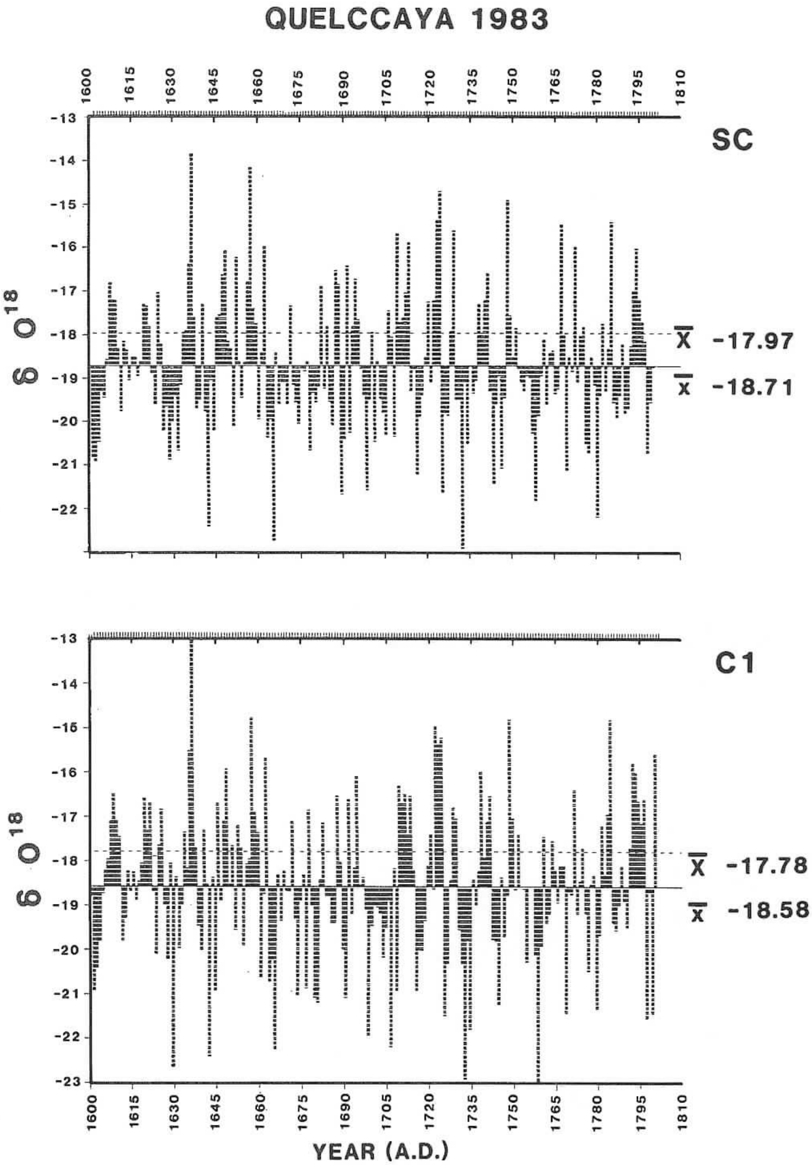

Figure 27.9 Identical to Figure 27.8, but representing the period from A.D. 1600 to 1800, except the small x̄ is the average for the 200-year period illustrated.

27.10 show that for both Quelccaya ice cores: (1) A.D. 1400 to 1530 was a period of average interannual isotopic variability; (2) around A.D. 1530 the annual averages of $\delta^{18}O$ become more negative and remain generally more negative until around A.D. 1880 and (3) the annual averages of $\delta^{18}O$ of the twentieth century are generally less negative. These annual oxygen isotopic values indicate a greater interannual variability during the LIA which is particularly pronounced in Figure 27.9 representing the period A.D. 1600 to 1800. Figures 27.11, 27.12

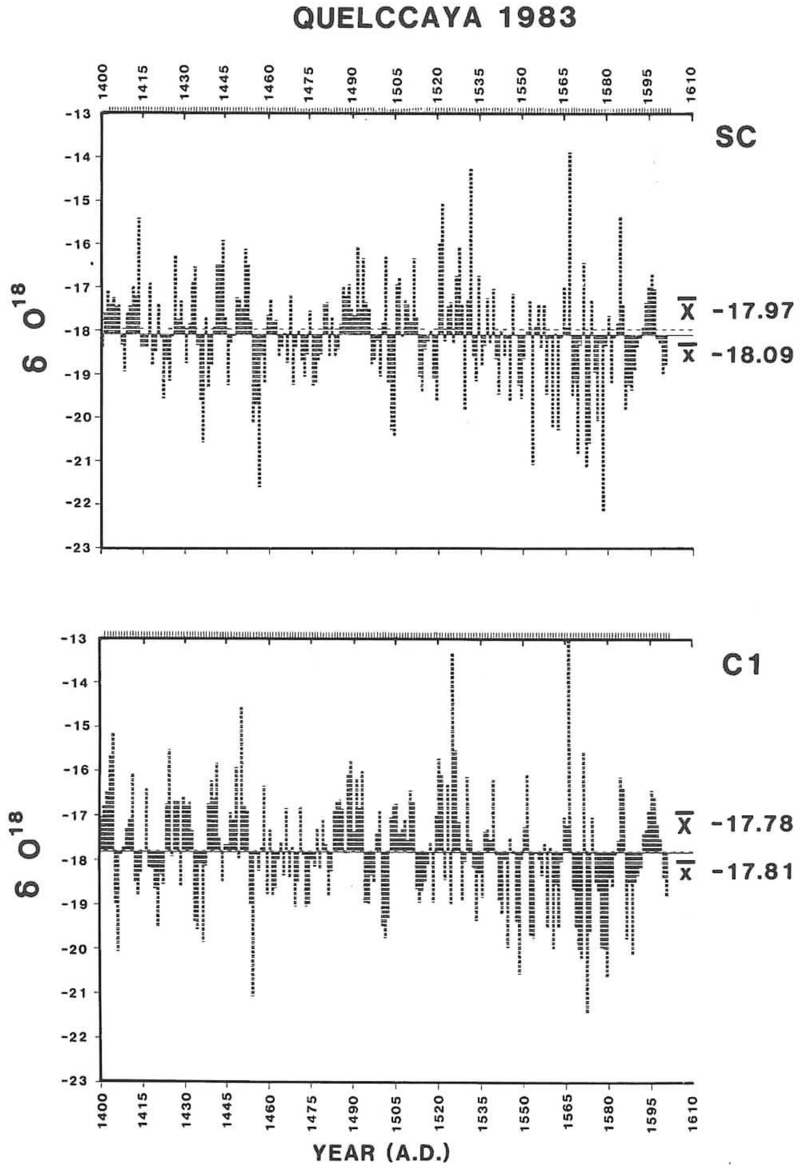

Figure 27.10 Identical to Figure 27.8, but representing the period from A.D. 1400 to 1600, except the small x̄ is the average for the 200-year period illustrated.

and 27.13 illustrate the interannual variability in total particles (insoluble dust). These data indicate a greater variability both on an interannual comparison and between cores throughout the A.D. 1400 to 1984 period than seen in either the accumulation or oxygen isotope records. Total particle concentrations increase slightly during the LIA period as can be seen in Table 27.2. These figures demonstrate that it is essential that spatial, as well as temporal, variability of the physical and chemical stratigraphic records be determined.

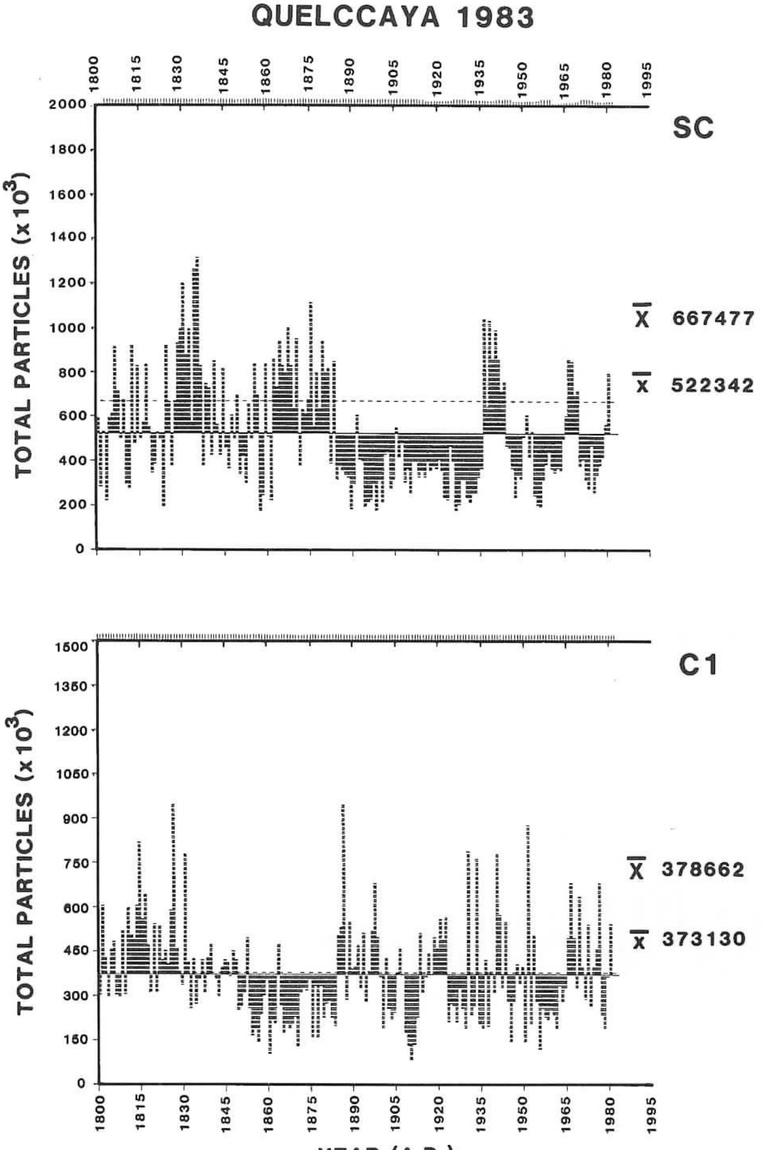

Figure 27.11 Mean annual particulate concentration (diameters from ≥ 0.63 and $\leq 16.0\mu m$ per ml) in the summit core (SC) and core 1 (C-1) for period A.D. 1800 to 1980. Large $\bar{X}$ is the average for the entire core, small $\bar{x}$ is the average for the 180 year period illustrated.

Figure 27.14 shows the annual records of total particles (diameter $>0.63\mu m$) conductivity, $\delta^{18}O$ and net accumulation for the last 500 years as reconstructed from the Quelccaya summit ice core. The A.D. 1600 eruption of Huaynaputina, the most explosive event recorded in the central Andes of Peru, is very distinct in the particulate and conductivity records (Thompson *et al.* 1986). However, no marked increase in insoluble or soluble dust occurs in the 1815 to

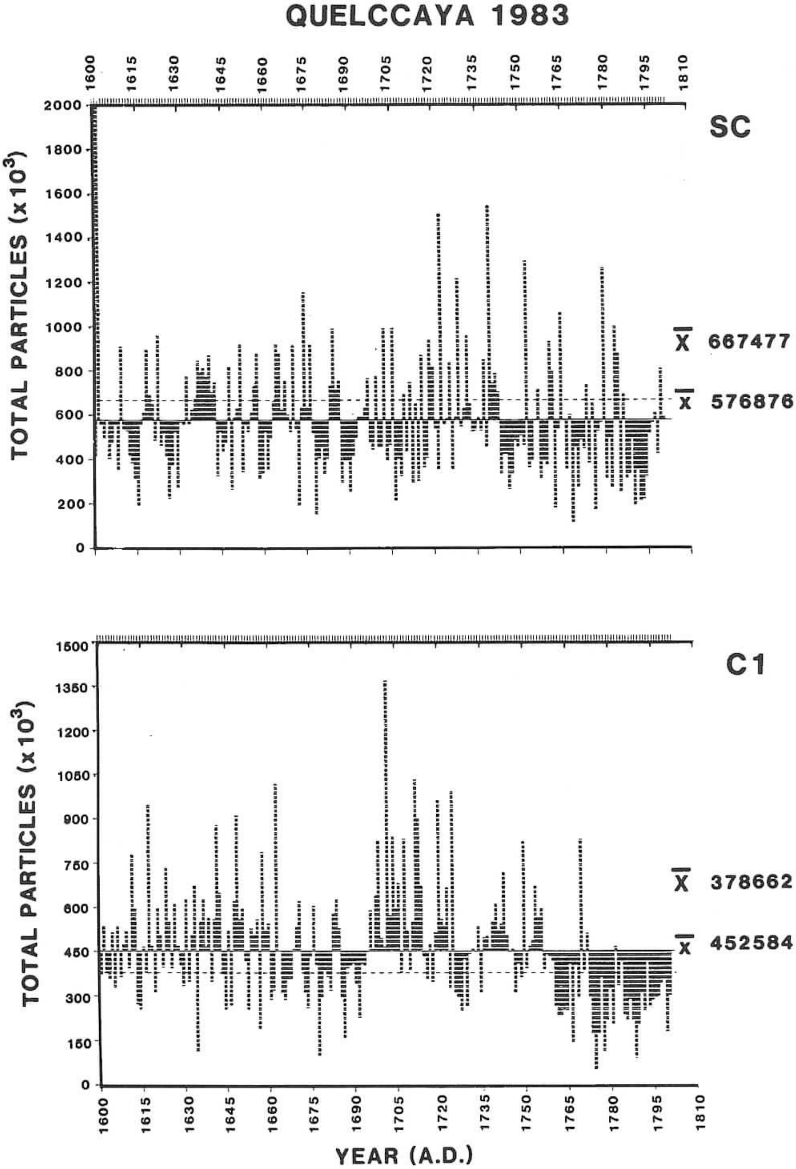

Figure 27.12 Identical to Figure 27.11, but representing the period from A.D. 1600 to 1800, except the small $\bar{x}$ is the average for the 200-year period illustrated.

1820 period associated with the 1815 eruption of Tambora. This is not unexpected in view of predominance of the local Altiplano as the major dust source which may mask more distant dust events. The $\delta^{18}O$ record, however, shows a prominent cooling which reaches a minimum in A.D. 1819-1820. The most negative $\delta^{18}O$ values in the entire 1500-year record occur in the wet season snowfall of this year as illustrated in Figure 27.3.

During the LIA in southern Peru, A.D. 1490-1880, microparticles and conductivities were

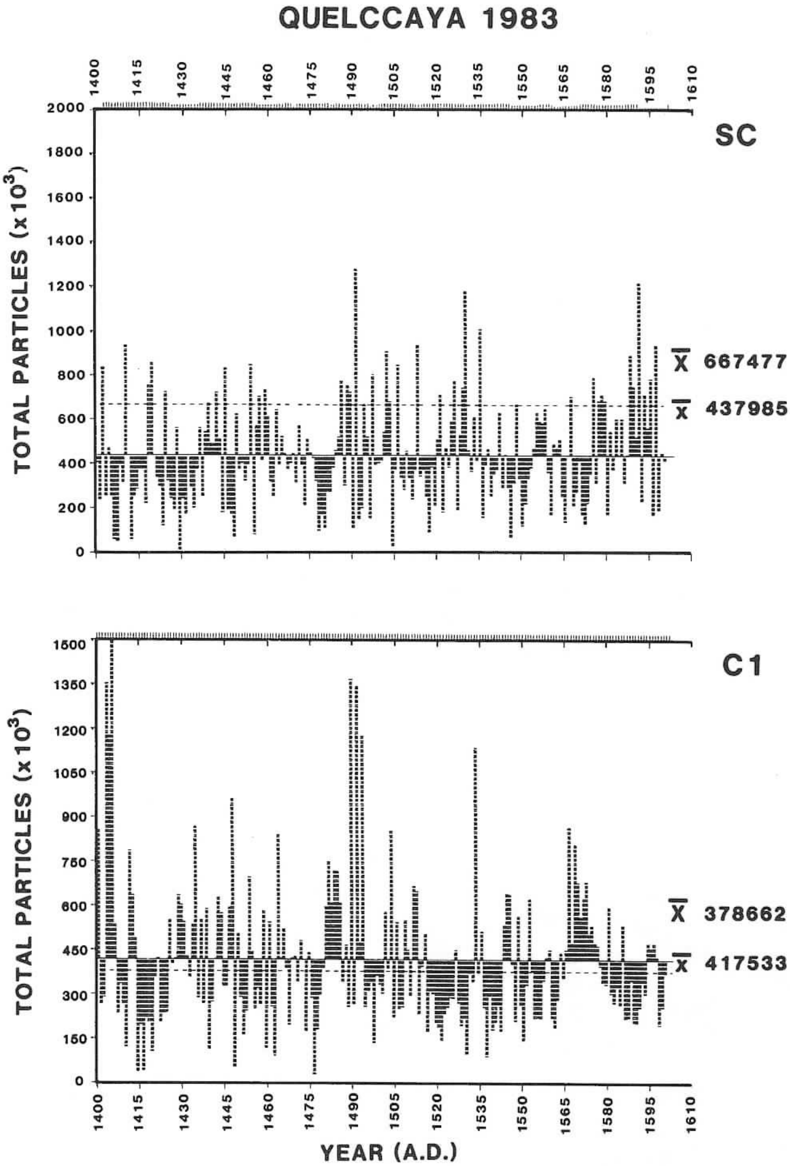

Figure 27.13 Identical to Figure 27.11, but representing the period from A.D. 1400 to 1600, except the small x̄ is the average for the 200-year period illustrated.

20 to 30% above the respective averages for the entire core. Increases in particulates may result from increased atmospheric impurities and/or decreased accumulation. The accumulation history (Figures 27.5, 27.6, 27.7 and 27.14) from Quelccaya is well documented which suggests that the increase in particulates must be due to increased atmospheric loading, not decreased accumulation. Figure 27.14 illustrates that the period A.D. 1500 to 1720 was the wettest interval in the 600 years. Similarly, A.D. 1720 to 1860 was very dry and yet both

Table 27.2 Quelccaya summit and Core 1 data summary.

Time (A.D.)	$\delta^{18}O$ (‰)		Averages ($\bar{x}$) and Differences (Δ) Total Particles ≥ 0.63 to ≤ 16.0 μm in diameter per ml of sample		Conductivity (μS cm-1)	
			Summit Core			
	$\bar{x}$	(Δ)	$\bar{x}$	(Δ)	$\bar{x}$	(Δ)
1880-1983	-17.69		430,000		1.95	
		0.98		158,000		0.85
1530-1880	-18.67		588,000		2.80	
		0.85		105,000		0.63
1250-1530	-17.82		483,000		2.17	
% change: 1880-1983 / 1530-1880	5.3%		26.8%		30.4%	
% change: 1250-1530 / 1530-1880	4.6%		17.9%		22.5%	
			Core 1			
	$\bar{x}$	(Δ)	$\bar{x}$	(Δ)		
1880-1983	-17.85		373,000			
		0.66		47,000		
1530-1880	-18.51		420,000			
		0.65		31,000		
1250-1530	-17.86		389,000			
% change: 1880-1983 / 1530-1880	3.6%		11.2%			
% change: 1250-1530 / 1530-1880	3.5%		7.4%			

particulates and conductivities remain unchanged from previous wet period values. Thus, the variations in microparticles and conductivity concentrations cannot be explained simply by changes in the rate of snow accumulation. Preliminary SEM and light microscope analyses of insoluble particles show no significant changes in the types of particles deposited during the LIA. Therefore, it is most likely that increased particulate deposition resulted from increased wind velocities across the high altiplano of southern Peru. Additional support for this interpretation is found in that the number of large particles (diameter > 1.59 μm) also increase during the LIA period (Thompson *et al.* 1986). Between A.D. 1530 and 1880 the

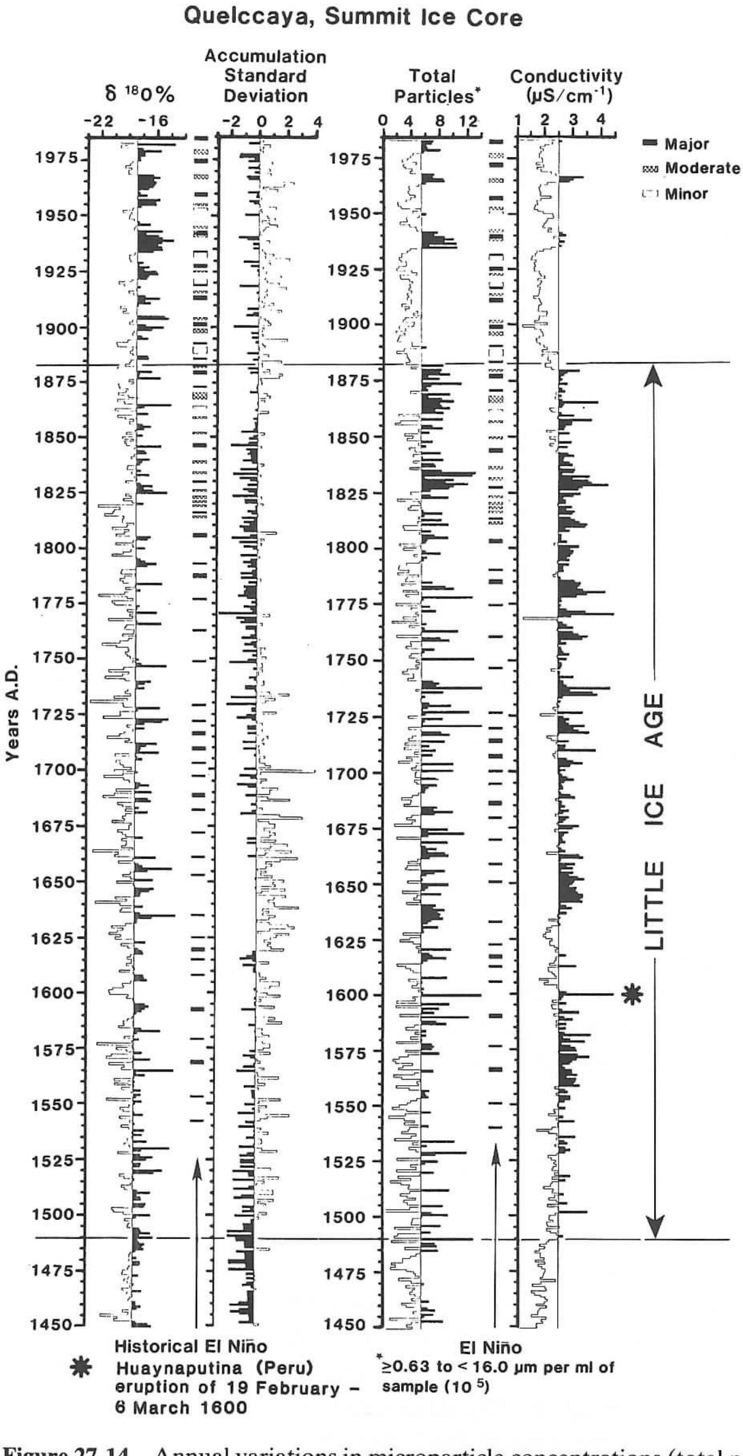

Figure 27.14 Annual variations in microparticle concentrations (total particles) conductivity, oxygen isotope ratios and accumulation (in standard deviation units, based on the last 500 years). 1(sigma) is equivalent to 34cm of accumulation. The "Little Ice Age" stands out clearly and is characterized by increased soluble and insoluble dust and decreased (more negative) $\delta^{18}O$. It appears to have been a major climate event in tropical South America. The large dust event centered on A.D. 1600 was produced by the February 19-March 6 A.D. 1600 eruption of Huaynaputina, Peru. Also shown is the historical El Niño record from Quinn *et al.*, 1987 (cf. Quinn and Neale, this volume).

average oxygen isotope ratios are 0.9‰ lower than in preceding or subsequent periods (see Table 27.2).

27.5 Abrupt climatic changes

The total ecosystem, and particularly the future well-being of man, is clearly as affected by the "rate of climatic change" as by the magnitude of climate change which actually occurs. In general, abrupt changes in climate denote a rupture of the established range of experience, and would be an unexpected and surprisingly fast transition from one state to another (Berger and Labeyrie 1987). Recent research (Dansgaard *et al.* 1989; Jouzel *et al.* 1987; Oeschger *et al.* 1984; Thompson and Mosley-Thompson 1987, 1989b) demonstrates the variety of parameters which can be measured for these periods and how ice core records can be used to define very precisely the rate of climate change.

These abrupt changes seem to occur on a spectrum of time scales. For example, in Greenland ice cores the transition from the last glacial to the present interglacial (~10,000 years ago) occurred in less than a hundred years (Thompson 1977) or perhaps in less than thirty years (Herron and Langway 1985) while the termination of the Younger Dryas climate event may have occurred in less than twenty years (Dansgaard *et al.* 1989). In the Dunde ice cores (China) the glacial-interglacial transition in dust concentration is completed within a 30cm section of ice representing approximately forty years (Thompson *et al.* 1989a; Thompson *et al.* 1990). These terminations may reflect steep gradients across the site in question, such as frontal zones in the climatic system, however, when they occur at the same time at many points on the globe, then they must reflect abrupt changes in the climate system which are not just site specific.

The annual record of climate variations for the last 500 years from the upper portion of the 1500-year Quelccaya, Peru, ice core record (shown in Figure 27.14) shows abrupt changes in tropical South America (Thompson and Mosley-Thompson 1987; 1989) with the transition from the LIA to the warmth of the current century occurring over a two to three year period centered on A.D. 1880. It was marked by a sharp change from the high amplitude seasonal oxygen isotope variations which characterized the Little Ice Age (LIA) period (Figure 27.15). The seasonal $\delta^{18}O$ range which averages 2‰ for the period A.D. 1880 to 1980 is twice as large, averaging 4‰, for the LIA period (A.D. 1520 to 1880) with a maximum of 13‰ in A.D. 1819-1820. Both cores illustrate a period during the LIA from A.D. 1590 to 1630 when the large seasonal $\delta^{18}O$ oscillations abruptly decrease. The large annual range in $\delta^{18}O$ during the LIA may reflect increased seasonality during this period, a feature noted in the historical records of Europe (Gribbin and Lamb 1978; Lamb 1984a, b). Alternatively, the annual signal may have been better preserved under the climatic conditions which existed during much of the LIA period. Regardless of the cause, the change in ice core records clearly illustrates that rapid alternation of the climate or environmental conditions occur around A.D. 1520, 1590, 1630 and at the termination of the LIA in A.D. 1880. Thus, during the LIA not only was there an apparent increase in the interannual variability as seen in Figures 27.8, 27.9 and 27.10, but also an increase in intrannual variability as seen in Figure 27.15.

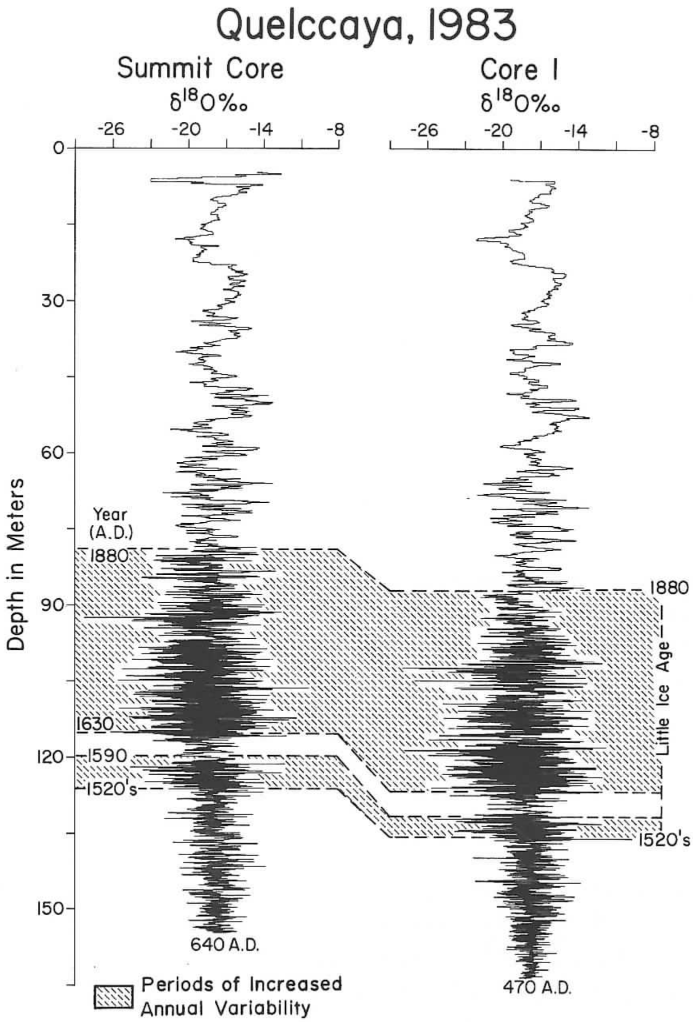

Figure 27.15 The annual range in oxygen isotope values in the Quelccaya summit core and core 1. The annual range in $\delta^{18}O$ increases by 100 percent during the Little Ice Age, reaching the largest values between A.D. 1650 to 1710. Low ranges occur in both pre- and post-Little Ice Age periods as well as in the unusual period from A.D. 1590 to 1630. Periods of high oscillations are shaded.

27.6 Comparison of Quelccaya and Dunde Ice core records

The Qinghai-Tibetan Plateau is a major heat source for the lower atmosphere and thus, greatly affects the hemispheric circulation. The present-day northern summer-monsoon circulation is driven by the relative warmth of the African-Asian landmass in comparison with the cooler surrounding ocean (Fein and Stephens 1987).

The interannual fluctuations in the summer monsoon rains over southeastern Asia have a profound socio-economic impact on China and India. Weak monsoons are associated with drought, crop failure, and, in extreme cases, famine. Strong monsoons are associated with devastating floods and the accompanying loss of life, property and crops. The importance of the Qinghai-Tibetan Plateau as a heat source has been noted by many authors (e.g., Flohn 1957, 1965, 1968; Koteswaram 1958; Murakami 1958, 1981; Virji 1979; Yeh 1981; Luo and Yanai 1983; Reiter 1983; Lau and Li 1984; Barnett *et al.* 1988; Hansen *et al.* 1988).

Eastern Asia is directly influenced by the thermal and dynamical forcing enhanced by this huge elevated landmass. Sensible heat flux over the semi-arid western region of the Plateau and latent heat release above the Himalayas, contribute to a strong tropospheric heat source which maintains the large-scale Asian monsoon circulation. Lau and Li (1984) point out that not only does the Plateau determine the large-scale circulation, but also strongly influences synoptic scale monsoon events over China. Unfortunately, there are no long historical records of climatic variability from the far western sections of the Plateau. However, the detailed historical records of drought, flood and dust storms from central and eastern China provide independent ties for assessing the spatial significance of events on the Plateau.

The present-day glaciers in China are widely scattered, mainly across the Qinghai-Tibetan Plateau, and cover a large area of about 57,000km^2 (Shi and Wang 1979). The snow line elevations range from 3000m in Altay Shan in northwestern China, to 6200m in southern and western areas of the Plateau. The annual precipitation ranges between 0.8 to 2.5m in parts of the outer mountains, and declines to only 0.2 to 0.3m on the inner Plateau. The greater absorption of solar radiation over the vast expanse of the Plateau than on the surrounding lowlands also raises the snowline elevations.

From about A.D. 1400 to 1900, temperatures in China were 1 to 2°C lower than present and resulted in prominent advances of glaciers. According to Zhu Kezhen (Chu Ko-Chen) (1973) the lowest temperatures of the last 500 years in central and eastern China occurred around A.D. 1700, clearly within the LIA. The Dunde oxygen isotope records reveal a cold period in western China from A.D. 1670 to 1740. The comparison of the Dunde ice cap decadal average $\delta^{18}O$ from A.D. 1580 to 1980 with Quelccaya (Figure 27.16) shows a general correspondence to Landsberg's first approximation for decadal Northern Hemisphere temperature departures for this period (Landsberg 1985; Groveman and Landsberg 1979). The correlation coefficients between decadal oxygen isotope records from Dunde and Quelccaya records and Landsberg's northern hemisphere temperature departures are presented in Table 27.3. All ice core records in Figure 27.16 show isotopic values below the A.D. 1880-1980 mean for most of the LIA, while all of the records illustrate the warming of the twentieth century.

The study of climate change on the time scale of centuries to millennia provides a valuable time perspective for interpreting climatic patterns of the present century and assessing the potential future effects of climate change due to a greenhouse warming. For example, there has been considerable debate regarding when and if a "greenhouse" signal will rise above the "climate noise" or variance within the unperturbed climatic system on time scales of decades to centuries. However, to make more confident projections about future climate change, it is necessary first to describe and understand the sources of that variance on the same time scale (Zhang and Crowley 1989).

In the monsoon region of China the precipitation has a marked annual cycle with 80% of the rain or snow (higher elevation sites) generally occurring in the summer half-season. The

Table 27.3 Statistical correlations (r) among the decadal averages of $\delta^{18}O$ for Dunde and Quelccaya ice caps and Landsberg's decadal temperature departures. Deviations are calculated with respect to the 1881 to 1980 mean. r is the linear correlation coefficient, r^2 is the coefficient of determination, and p is the level of statistical significance.

	Quelccaya Summit Core $\delta^{18}O$	Northern Hemisphere Temperature Departures
Dunde Core D-1 $\delta^{18}O$	r = 0.433 r^2 = 18.7% p = 99.5%	r = 0.482 r^2 = 23.2% p = 99.8%
Northern Hemisphere Temperature Departures	r = 0.632 r^2 = 39.9% p = 99.9%	

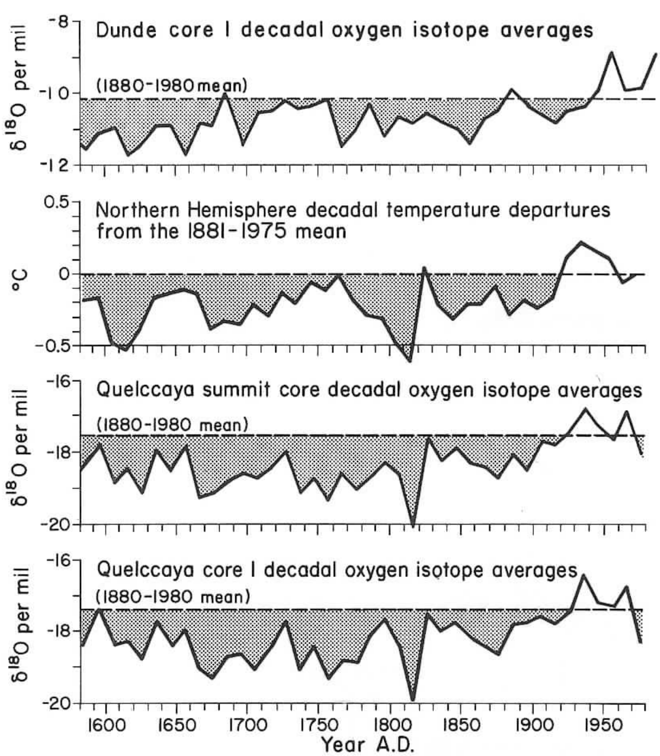

Figure 27.16 Decadal temperature departures (from 1881-1975 mean) in the Northern Hemisphere from A.D. 1580 to 1975 compared with decadal average $\delta^{18}O$ values for both Quelccaya ice cores and the Dunde D-1 core. The dashed line is the A.D. 1881-1980 mean for the $\delta^{18}O$ records.

rainy season is closely associated with the position of the polar front, which is generally located at the northern periphery of the subtropical high. The retreat of the polar front accompanies the advance of the summer monsoon. The intensification of the Hadley circulation is a result of intense summer heating on the Eurasian landmass (e.g., Lau and Li 1984).

Lau *et al.* (1988) summarize the development of the summer monsoon in three stages: (1) the late spring position of the rainbelt is in south China, and its early summer (mainly June and early July) position is in the Yangtze Valley (the so-called *Mei-Yu* or "Plum Rains"); (2) the northernmost positions reach China and inner Mongolia in July-August; (3) in autumn the rainbelt retreats to the south and persistent rain is prominent only in southwest China, although typhoon rains peak at this time and can significantly influence rainfall totals. In winter the whole country is dry except in the Yangtze Valley, where protracted small rains are frequent and under the influence of a (subtropical) southwesterly jet stream.

Figure 27.17 illustrates the initial comparison of decade-averaged accumulation in meters per year for the period A.D. 1610 to 1980 for Quelccaya and Dunde ice core records. These records show a general similarity in net balance for the last 400 years for these two widely separated areas. The correlation between the decadal averaged net accumulation for Dunde and Quelccaya increase from $r = 0.29$ with no lag to $r = 0.57$ (significance = 99%) when Dunde record is lagged 50 years. It has been recognized since the turn of the century that teleconnections link these two widely separated sites. The El Niño-Southern Oscillation

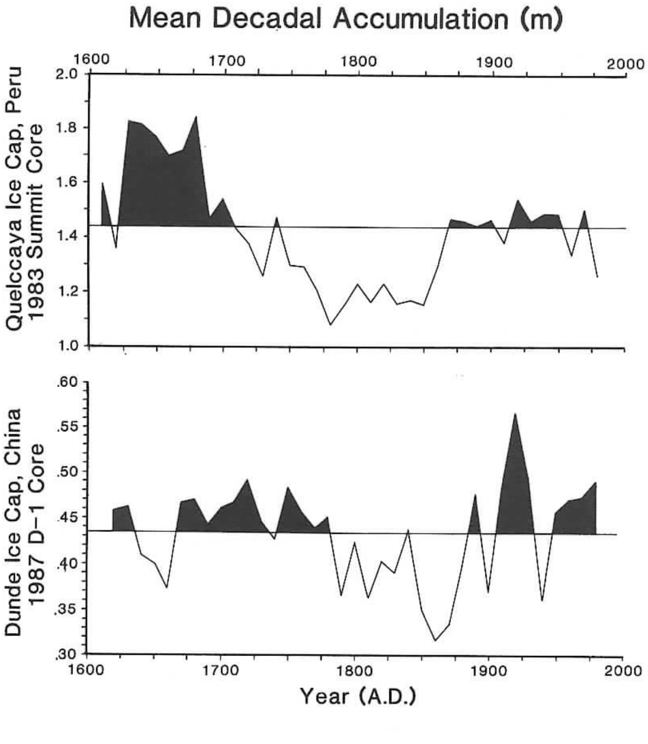

Figure 27.17 Decadal accumulation (net balance) in meters for the Quelccaya, Peru summit core compared with the Dunde, China core D-1 for the period A.D. 1610 to 1980.

phenomenon perturbs the ocean-climate system of the Pacific basin episodically and is related to anomalous weather patterns covering much of the globe (Rasmussen and Wallace 1983; Nicholls 1987; Enfield 1989). The longer-term similarity of patterns in accumulation illustrated in Figure 27.17 suggest that teleconnections may exist between these widely separated areas not only for high frequency events such as El Niño-Southern Oscillation phenomenon, but also for low frequency events lasting centuries. Further evidence for large scale teleconnections are also suggested by the comparison of the Quelccaya and Dunde ice cap accumulation records with those from Mt.Logan, Canada (Holdsworth *et al.*, this volume).

27.7 Climate, man and the environment in South America

The historical record of man's activities in pre-Spanish South America is scant and incomplete, and the process of piecing it together is hampered by the lack of written language. Archaeologists, through excavations of sites of pre-Hispanic civilizations in Ecuador and Peru, have been able to better define the chronology and activities of civilizations in these regions. Additional pieces of the historical puzzle can be provided by non-archeological sources. For example, the Quelccaya ice cores have provided records of climatic events which may have played a major role in prehistoric man's activities.

The archeological sites in Peru may be assigned to either coastal or highland cultures. The accumulation (meters of ice) record for the southern Peruvian highlands presented in Figure 27. 18 is a composite of the decadal averages from both core 1 and summit core. Four very marked dry periods are found: A.D. 540 to 610, A.D. 650 to 730, A.D. 1040 to 1490 and A.D. 1720 to 1860, and two distinct wet periods occur from A.D. 760 to 1040 and A.D. 1500 to 1720. It is important to note that under the present climate regime, coastal Ecuador and coastal Peru suffer from heavy rains in association with El Niño-Southern Oscillation events (ENSO) while the highlands of southern Peru, where the Quelccaya ice cap is located, often experience drought conditions (Thompson *et al.* 1984b; Lam and Del Carmen 1986). Paulsen (1976) reported a similar relationship in the longer term archeological records establishing the rise and fall of coastal cultures of Peru and Ecuador. For comparison, the cultural record (dating based largely on highly refined ceramic sequences and some ^{14}C measurements) from Peru and Ecuador are also presented in Figure 27. 18. Paulsen reported that highland and coastal cultures seemed to flourish out of phase with each other, i.e., highland cultures flourish when coastal cultures decline and *vice versa*. Since both the prehistoric coastal and highland civilizations were largely agrarian and both the coastal areas (due to dependence on a limited water supply) as well as the high plateau areas (being the upper limits of agriculture) are very climatically sensitive (Cardich 1985) it is certain that climate played an important role in the survival of these cultures. However, the exact nature of that role of climatic variability as a dominant independent variable in prehistoric Andean cultural changes is debated (Stahl 1984; Paulsen 1984).

The comparison of the records of accumulation from the Quelccaya ice cores and the archeological history demonstrate that highland cultures flourished during wet periods on the Bolivian-Peruvian plateau and conversely, coastal cultures flourished when the highlands were dryer. Assuming that the same seesaw relationship which currently exists during ENSO

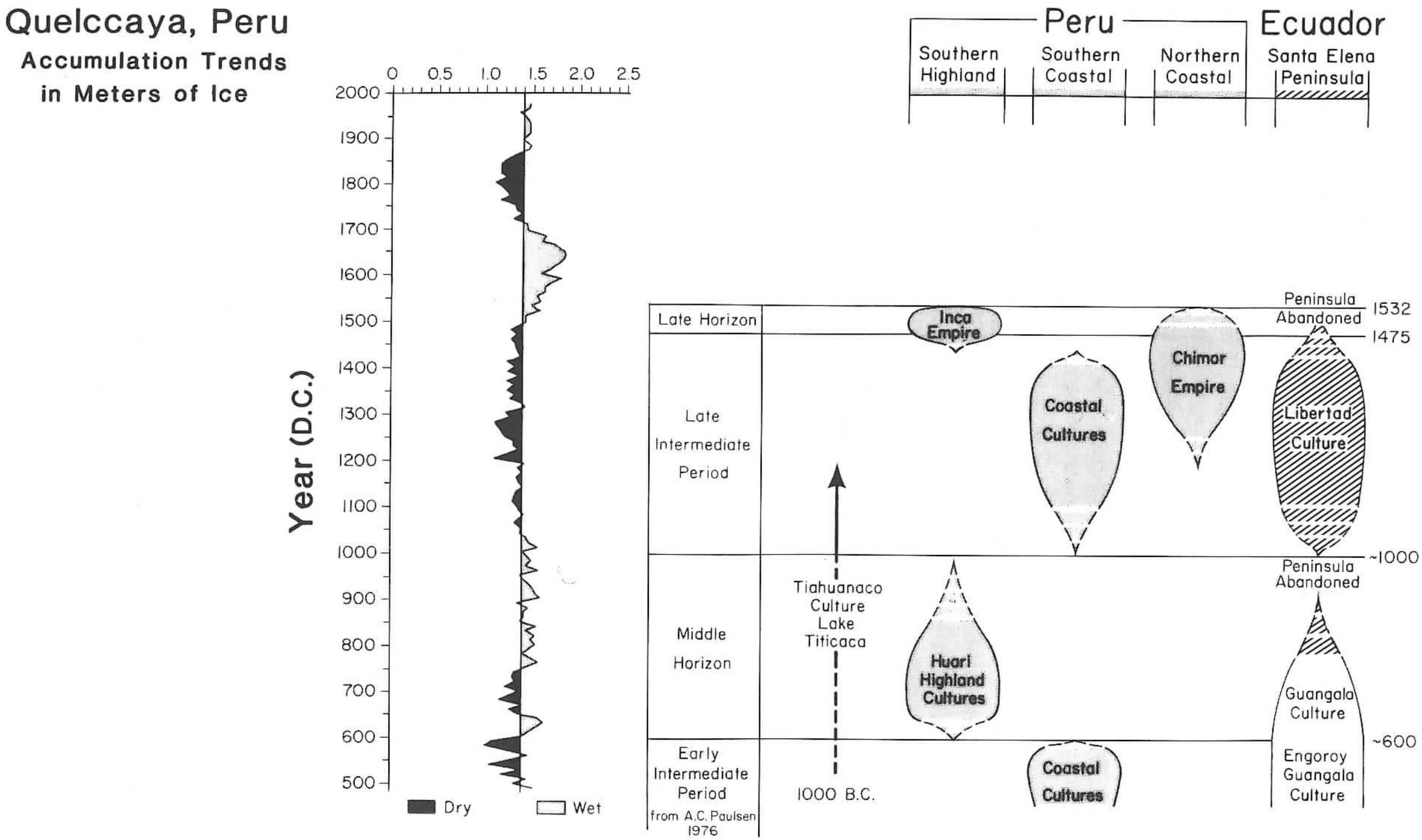

Figure 27.18 Quelccaya decadal accumulation trends in meters of ice are presented as a composite of core 1 and summit core records. Wet and dry periods are noted. On the right the periods of the rise and fall of coastal and highland cultures of Ecuador and Peru are illustrated (taken mainly from Paulsen, 1976).

events (Thompson *et al.* 1984b) was valid over the longer periods of El Niño-like ocean and atmospheric circulation patterns, then wetter coastal conditions would be expected during periods of highland drought. The fact that coastal cultures flourished during the dryer highland periods implies that this seesaw relationship existed over extended time periods.

Correlations have been established between the shifts of cultural activity between highland and coastal civilizations in the region and the accumulation record obtained from the ice cores. This record may also provide a key to reconstructing the succession of habitation and abandonment of sites of so called "lost" civilizations uncovered by archaeologists, such as the Gran Pajaten. However, a correlation between past environmental events and cultural events does not necessarily imply a direct connection. When there is a connection, it involves that culture's adaptation to new conditions and the mode of the adaptation depends on the culture.

27.8 The global perspective

The assimilation of multifaceted ice core data sets from polar and non-polar regions makes possible the determination of both temporal and geographical variability of any or all the individual ice core parameters, i.e., dust, chemistry, gas content, isotopes, etc. The $\delta^{18}O$ records are discussed on a global scale using examples in Figure 27.19. The $\delta^{18}O$ records reflect in varying degrees (1) air temperature at which condensation takes place, (2) atmospheric processes between the oceanic source of water vapor and the site of deposition, (3) local conditions under which the isotope signal is modified during firnification, (4) the surface elevation of the depositional site and (5) the latitude of the site (see Dansgaard *et al.* 1973 and Bradley 1985, for review). Although the correlation of atmospheric temperatures with oxygen isotope ratios and its spatial representativeness is still under discussion, the method continues to be used widely as a proxy for climate and in particular for temperature (Dansgaard *et al.* 1973; Jouzel *et al.* 1983; Thompson *et al.* 1986 and Peel *et al.* 1988). Figure 27.19 provides a preliminary global perspective of the decadal and centennial variations for the periods ranging between 550 to 1000 years B.P. from five sites across the globe. The sites are, from north to south, Camp Century, Greenland (Johnsen *et al.* 1970); Dunde ice cap, China (this paper); Quelccaya ice cap, Peru (Thompson *et al.* 1986) and South Pole and Siple Station, Antarctica (Mosley-Thompson *et al.*, in press). In Figure 27.19, isotopically heavier, and hence warmer, periods are illustrated by shaded areas and isotopically lighter, and hence colder, periods are unshaded. The records show a large diversity in detail not unlike that which would be found if only five widely dispersed meteorological stations were used to reconstruct global temperatures. However, several features stand out, such as the similarity of variability of isotopes in the two northern hemisphere sites with a rather pronounced 180-year oscillation in the China decadal oxygen isotopic record. In the southern hemisphere there is a strong similarity in the oxygen isotopic records from the tropical Quelccaya ice cap, Peru and South Pole, Antarctica. The LIA stands out as a period of more negative isotopic values from ~A.D. 1530 to 1900, while before and after, both locations were generally isotopically less negative. Of these five records, the Siple Station record is unique because during the LIA this site exhibits less negative oxygen isotopes and hence presumably warmer conditions (see Mosley-Thompson, this volume). All sites show isotopically less negative

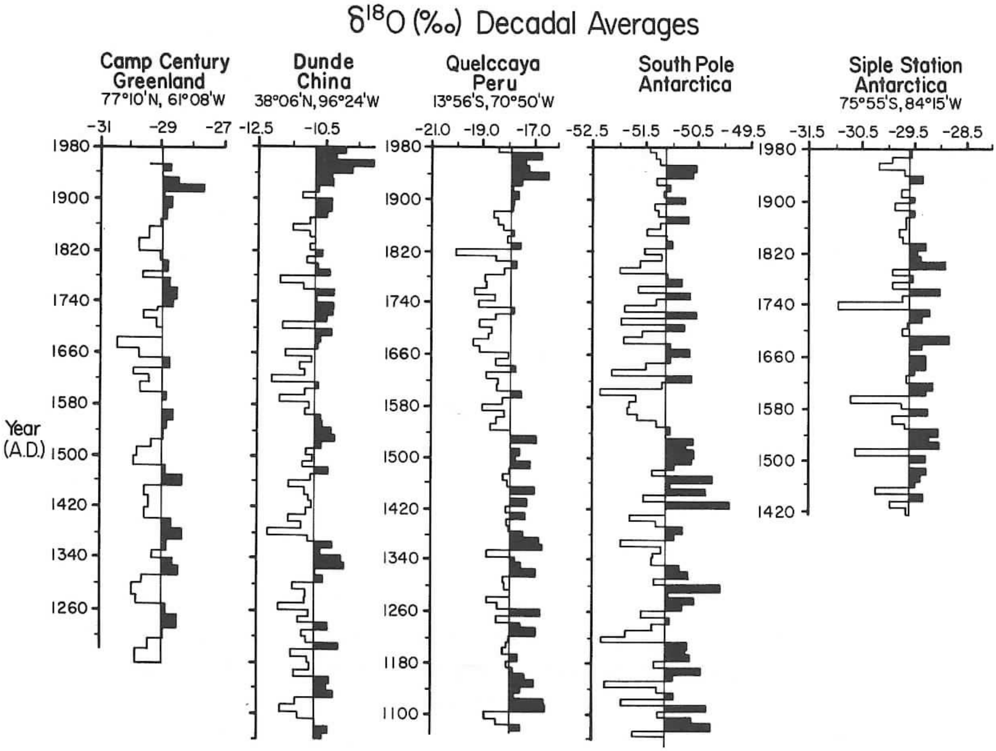

Figure 27.19 Decadal averages of the $\delta^{18}O$ records in a north-south global transect from Camp Century, Greenland to the South Pole, Antarctica. The shaded areas represent isotopically less negative, or warmer, periods and the unshaded areas represent isotopically more negative, or colder, relative to the respective means of the individual records.

conditions during the twentieth century, except Siple Station which is, in general, isotopically more negative for the last 100 years. However, for the last thirty years the isotopic evidence (Peel *et al.* 1988; Mosley-Thompson *et al.*, in press) indicates a general trend to less negative values and thus warmer conditions which is consistent with the warmer overall, atmospherically-measured temperatures in the Antarctic Peninsula from 1960-1980.

A most striking feature of the $\delta^{18}O$ records in Figure 27.19 is the extreme warmth in central China during the last 60 years, with the warmest decades being the 1940s, 1950s, and 1980s. Using the $\delta^{18}O$ record as a proxy for temperature, the last 60 years constitute one of the warmest periods in the entire record, equalling levels of the Holocene maximum 6000 to 7000 yr B.P. (Thompson *et al.* 1989a,; Thompson *et al.*, in press). Model results of Hansen *et al.* (1988) suggest that the central part of the Asian continent may be one of the first places to exhibit an unambiguous signal of the anticipated "greenhouse warming." Certainly the Dunde ice core results suggest that the recent warming on the Tibetan Plateau has been substantial. Recent radiosonde data from Southern India (Flohn and Kapala 1989) show that, in fact, the average tropospheric temperature has increased nearly 1°C since 1965. It must be cautioned that more robust temperature-isotope transfer functions must be developed for the Tibetan Plateau and indeed for all regions of the earth where isotopic ice core records are being used as proxy indicators of temperature.

In conclusion, the ice core climatic and environmental records from low, middle and high latitudes can greatly increase our knowledge and understanding of the course of climatic events of the past. This is essential to predict future climatic oscillations, which may or may not be dominated by increasing greenhouse gas concentrations. The forcing factors, internal and external, which have operated in the past to prevent climatic stability will continue to operate and influence the course of events (Grove 1988). The cores from the Dunde ice cap, China cores provide the first ice core record of the Holocene/late Pleistocene climate from the subtropics. The stable isotope record indicates that the last 60 years on the Tibetan Plateau constitutes one of the warmest periods in the entire record. Events such as the LIA are global in scale and thus result from large scale climatic forcing which influences the entire earth system. The manifestation of the LIA in any given record is quite variable and more distinct in the higher elevation sites, such as Dunde, Quelccaya and South Pole, than in the lower elevation sites of Camp Century and Siple Station. This may imply importance of climatic change as a function of elevation as well as of the more subtle changes in climate like the LIA. The latter are more clearly recorded farther from the mitigating influence of the oceans. The high temporal resolution available from ice cores illustrates that the transition from climatic "norms" may be abrupt on the scale of the LIA and glacial/interglacial events. The tropical and subtropical ice core records provide the potential to establish long records of El Niño-Southern Oscillation events and monsoonal variability. Information of variations in their magnitude and frequency through time potentially may provide information on the causes of these global-scale events. Moreover, a global array of cores can be used to establish changes in precipitation over the last 1000 to 2000 years.

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