CLIMATIC CHANGE AT HIGH ELEVATION SITES: AN OVERVIEW

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Abstract. This paper provides an overview of climatic changes that have been observed during the past century at certain high-elevation sites, and changes in a more distant past documented by a variety of climate-sensitive environmental indicators, such as tree-rings and alpine glaciers, that serve as a measure of the natural variability of climate in mountains over longer time scales.

Detailed studies such as those found in this special issue of *Climatic Change*, as well as those noted in this review, for the mountain regions of the world, advance our understanding in a variety of ways. They are not only helpful to characterize present and past climatological features in the mountainous zones, but they also provide useful information to the climate modeling community. Because of the expected refinements in the physical parameterizations of climate models in coming years, and the probable increase in the spatial resolution of GCMs, the use of appropriate data from high elevation sites will become of increasing importance for model initialization, verification, and intercomparison purposes. The necessity of accurate projections of climate change is paramount to assessing the likely impacts of climate change on mountain biodiversity, hydrology and cryosphere, and on the numerous economic activities which take place in these regions.

1. Introduction

Mountain systems cover about one-fifth of the earth's continental areas and are all inhabited to a greater or lesser extent except for Antarctica. Mountains provide direct life support for close to 10% of the world's population, and indirectly to over half. Because of their great altitudinal range, mountains such as the Himalayas, the Rockies, the Andes, and the Alps, exhibit, within short horizontal distances, climatic regimes which are similar to those of widely separated latitudinal belts; they consequently feature high biodiversity. Indeed, there is such a close link between mountain vegetation and climate that vegetation belt typology has been extensively used to define climatic zones and their altitudinal and latitudinal transitions (cf. for example Klötzli, 1984, 1991, 1994; Ozenda, 1985; Quezel and Barbero, 1990; Rameau et al., 1993).

Mountains are also a key element of the hydrological cycle, being the source of many of the world's major river systems. Shifts in climatic regimes, particularly precipitation, in space or seasonally in a changing global climate, would impact heavily on the river systems originating in mountain areas, leading to disruptions of the existing socio-economic structures of populations living within the mountains and those living downstream.

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Although mountains may appear to be mere roughness elements on the earth's surface, they are in fact an important element of the climate system. They are one of the trigger mechanisms of cyclogenesis in mid latitudes, through their perturbations of large-scale atmospheric flow patterns. The effects of large-scale orography on the atmospheric circulation and climate in general have been the focus of numerous investigations, such as those of Bolin (1950), Kutzbach (1967), Manabe and Terpstra (1974), Smith (1979), Held (1983), Jacqmin and Lindzen (1985), Nigam et al. (1988), Broccoli and Manabe (1992) and others. One general conclusion from these comprehensive studies is that orography, in addition to thermal land-sea contrasts, is the main shaping factor for the stationary planetary waves of the winter troposphere in particular. The seasonal blocking episodes experienced in many regions of the world, with large associated anomalies in temperature and precipitation, are also closely linked to the presence of mountains.

A precise understanding of the climatic characteristics of mountain regions is complicated on the one hand by a lack of observational data at the spatial and temporal resolution adequate for climate research in regions of complex topography, and on the other by the considerable difficulty in representing complex terrain in current general circulation climate models (GCMs). Mountains are important perturbation factors to large-scale atmospheric flows; they also have an influence on the formation of clouds and precipitation in their vicinity, which are in turn indirect mechanisms of heat and moisture transfer in the vertical. Consequently, the influence of orography on climate needs to be taken into account in a physically-meaningful manner. Parametric schemes in GCMs take a number of possible forms, such as 'envelope topography' which smooths the real orography over continental areas. Large-scale orographic forcing is then derived by filtering the orography at smaller scales and effects of the atmospheric boundary layer. Gravity waves generated by the presence of the underlying orography on the atmosphere are capable of breaking in a similar manner to ocean waves at the seashore, and in doing so transfer substantial quantities of momentum from the large-scale to the small-scale flows.

Meteorological research has tended to focus on the upstream and downstream influences of barriers to flow and on orographic effects on weather systems (Smith, 1979), rather than on the specificities of climate within the mountain environments themselves. These include microclimatological processes which feed into the large-scale flows, and the feedbacks between the surface and the atmosphere, particularly vegetation and geomorphologic features, which can create microclimatic contrasts in surface heating, soil moisture or snow-cover duration (Geiger, 1965). Isolating macro-and microscale processes, in order to determine their relative importance, is complicated by inadequate data bases for most mountain areas of the world (Barry, 1994).

In terms of modeling studies of mountain climates, the dominant feature of mountains – i.e., topography – is so poorly resolved in most general circulation climate models (GCMs) that it is difficult to use GCM-based scenarios for investigating the potential impacts of climate change (Beniston et al., 1996). Any meaningful climate projection for mountain regions – and indeed for any area of less than a continental scale – needs to consider processes acting from the very local to the global scales. Numerous climatological details of mountains are overlooked by the climate models, making it difficult to predict the consequences of climate change on mountain hydrology, glaciers, or ecosystems (Giorgi and Mearns, 1991; Beniston, 1994). The situation is currently improving with the advent of high-resolution climate simulations, where the spatial scale of GCMs is on the order of 100 km (Beniston et al., 1995; Marinucci et al., 1995). However, much of the research on the potential impacts of climate change in mountain regions requires climatological information on scales which are generally far smaller than the typical grid-size of even the highest resolution numerical climate models. As a result, many impacts studies have been constrained by the lack of scenario data of sufficient reliability and quality at the desired scales. It is these scales, however, which are of particular interest to policy makers, especially in the context of the United Nations Framework Convention on Climate Change, whose aim is to limit anthropogenic interference with the climate system and allow its future evolution to be sufficiently slow for ecosystems to adapt naturally to this change.

The present paper will provide an overview of climatic change which has been observed this century at certain high-elevation sites, and changes in a more distant past as a measure of the natural variability of climate in mountains.

2. Particularities of Alpine Climate this Century

The climate of the Alpine region is characterized by a high degree of complexity, due to the interactions between the mountains and the general circulation of the atmosphere, which result in features, such as, gravity wave breaking, blocking episodes, and foehn winds. A further cause of complexity is to be found in the competing influences of a number of different climatological regimes in the region – Mediterranean, continental, Atlantic, and Polar. Traditionally, the European Alps are perhaps the best-endowed mountain system in terms of climatological and environmental data, and it is here that many of the most relevant studies of climate and climate change in mountain regions have been undertaken.

Figure 1 shows the changes in yearly mean surface temperature anomalies this century from the 1951–1980 climatological mean, averaged for eight sites in the Swiss Alps. These observational sites range in altitude from about 569 m above sea level (Zürich) to close to 2500 m (Säntis). The global data of Jones and Wigley (1990) has been superimposed here to illustrate the fact that the interannual variability in the Alps is more marked than on a global or hemispheric scale; the warming experienced since the early 1980s, while synchronous with the global warming, is of far greater amplitude and reaches up close to 1 °C for this ensemble average and up to 2 °C for individual sites, such as Säntis. This latter figure represents roughly

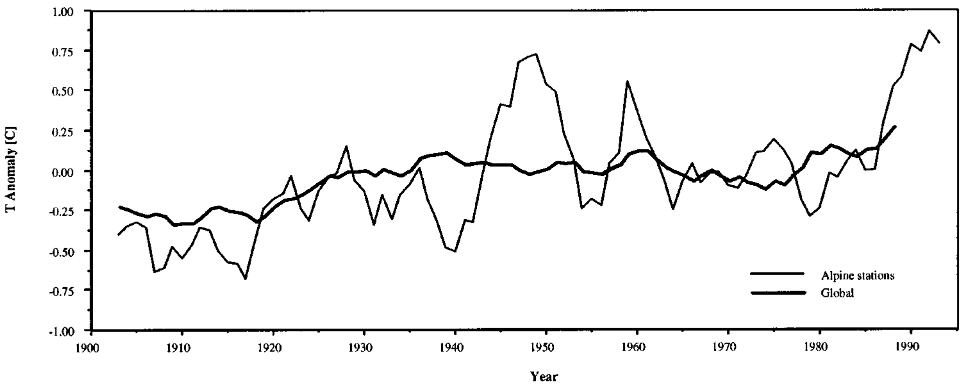

Figure 1. Yearly-mean surface temperature anomalies averaged for eight high-elevation sites in the Swiss Alps, ranging in altitude from 569 m to 2500 m above sea level. The change in global mean temperature anomalies is given for comparison purposes. Data have been smoothed with a five-year filter.

a five-fold amplification of the global climate signal (see also Diaz and Bradley, 1997). Similar studies for the Austrian Alps (e.g., Auer and Boehm, 1994) and the Bavarian Alps, based on climatological records from the Sönnblick Observatory (Austria) and the Zugspitze (Germany) lead to broadly similar conclusions.

Beniston et al. (1994) have undertaken an exhaustive study of climate trends this century in the Swiss Alps. Climate change in the region this century has been characterized by increases in minimum temperatures of about 2 °C, a more modest increase in maximum temperatures (in some instances a decrease of maxima in the latter part of the record), little trend in the precipitation data, and a general decrease of sunshine duration through to the mid-1980s. Warming has been most intense in the 1940s, followed by the 1980s; the cooling which intervened from the 1950s to the late 1970s was not sufficient to offset the warming in the middle of the century. The asymmetry observed between minimum and maximum temperature trends in the Alps is consistent with similar observations at both the sub-continental scales (e.g., Brown et al., 1992, for trends in the Colorado Rocky Mountains and neighboring Great Plains locations), and the global scale. Karl et al. (1993) have shown that over much of the continental land masses, minimum temperatures have risen at a rate three times faster than the maxima since the 1950s, the respective anomalies being on the order of 0.84 °C and 0.28 °C during this period.

Pressure statistics have been compiled as a means of providing a link between the regional-scale climatological variables and the synoptic, supra-regional scale. These statistics have shown that pressure exhibits a number of decadal-scale fluctuations, with the appearance of unusual behavior in the 1980s; in that particular decade, pressure reached annual average values far higher than at any other time this century. The frequency of occurrence of high pressure episodes (blocking high events) exceeding a sea-level reduced threshold of 1030 hPa between 1983 and 1992 account for one-quarter of all such episodes this century. The pressure field is well correlated with the North Atlantic Oscillation (NAO) Index – a measure of the strength of flow over the North Atlantic, based on the pressure difference between the Azores and Iceland – for distinct periods of the Swiss climate record (1931–1950 and 1971–1990) and is almost decorrelated from the NAO Index for the other decades of the century. This is indicative of a transition from one climatic regime to another, dominated by zonal flow when the correlation with the NAO Index is high. In the 1980s, when zonal flow over the North Atlantic was particularly strong, episodes of persistent, anomalously high pressures were observed over the Alps and in southern Europe, particularly during the winter season (see Hurrell and van Loon, 1997), as the eastern edge of the Azores high pressure cell extended systematically into south-western Europe. This resulted in large positive departures in temperature anomalies and a significant lack of precipitation, particularly in the form of snow, in the Alps. Indeed, much of the strong warming observed since the 1980s and illustrated in Figure 1 can be explained to a large degree by these persistent high pressure anomalies. During this same period in Europe, the Iberian Peninsula was under the influence of extended periods of drought and very mild winters, while northern Europe and Scandinavia were experiencing above-average precipitation (Hurrell, 1995).

Beniston and Rebetez (1996) have shown, that there is not only a time-dependency on observed warming, but also an altitudinal dependency of temperature anomalies. Taking the latter part of the climatological record into consideration, i.e., the 15-year period from 1979–1994, they have shown that minimum temperature anomalies exhibit a significant linear relationship with altitude, except at low elevations which are subject to wintertime fog or stratus conditions; the stratus or fog tends to decouple the underlying stations from processes occuring at higher altitudes. The authors have also shown that there is a switch in the gradient of the temperature anomaly with height from cold to warm winters. For warm winters, which represent 7 of the last 15 years, the higher the elevation, the stronger the positive anomaly; the reverse is true for cold winters (accounting for 5 winters during this period). This is illustrated in Figure 2 for the Alpine stations, which represent 39 of the 88 Swiss observational sites which were used in this investigation. The figure provides a measure of the damping of the climate signal as an inverse function of height. The spread of data is the result of site conditions; topography plays a key role in the distribution of temperature means and anomalies, in particular for stations located on valley floors. Here, thermal inversions are likely to occur which frequently lead to a decoupling of temperatures at these locations from those above the inversion and closer to free atmosphere conditions. Recent work by Giorgi et al. (1996), based on high-resolution simulations with a limited-area model have led to conclusions broadly consistent with the statistical analysis of Swiss data; an altitudinal dependency of temperature anomalies is perceived in GCM and regional climate model simulations (cf. Diaz and Bradley, 1997; Vinnikov et al., 1996).

One of the numerous problems associated with future climate change induced by anthropogenic emissions of greenhouse gases is the behavior of extreme temper-

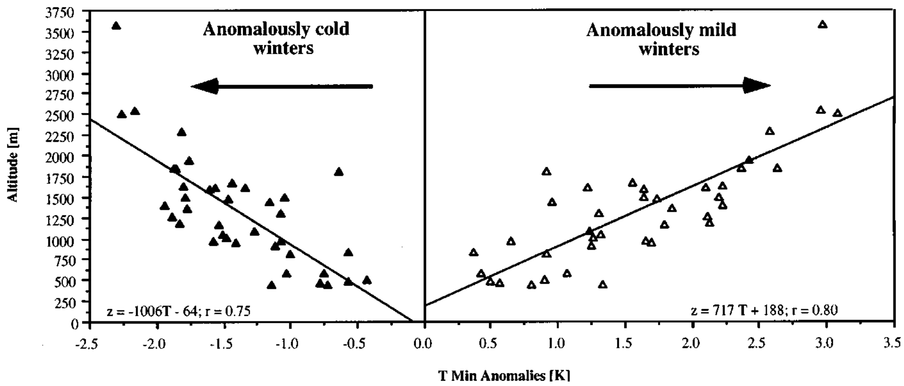

Figure 2. Vertical distribution of minimum temperature anomalies at 39 climatological stations in the Swiss Alps, for both 'mild' and 'cold' winters experienced during the 15-year period from 1979–1994.

atures and precipitation and how these may respond to shifts in the means (Giorgi and Mearns, 1991; Katz and Brown, 1992; Beniston, 1994; IPCC, 1996). Extreme climatological events tend to have a greater impact than changes in mean climate, which is commonly the basic 'quantity' for discussing climatic change, on a wide range of environmental and socio-economic systems. Based on the high spatial and temporal density of the Swiss climatological observing network, it has been possible to evaluate how shifts in climate this century have influenced the extremes of temperature (Beniston and Rebetez, 1997) at different locations in the Alpine region. Their study has shown that temperature extremes have been observed to shift by a factor of over 1.5 for a unit change in means this century; there is additionally an asymmetry between the shifts in the lower (minimum) extreme and the upper (maximum) extreme, leading to a changed frequency distribution profile for temperatures. This is illustrated in Figure 3 for the climatological station of Davos, located at 1,590 m above sea level in the eastern part of the Swiss Alps.

Precipitation in the warmer part of the climatological record (i.e., 1980–present) has been shown to exhibit a bimodal shift toward drier conditions, on the one hand, and more extreme precipitation events, on the other. While the relationships between temperature and precipitation extremes are difficult to quantify in a systematic manner, the above conclusion would imply that in a warmer global climate, precipitation amount in the Alps would be generally reduced, but isolated events of extreme precipitation could be expected to increase significantly. This is consistent with modeling studies of Schaer et al. (1996), who have determined that an increase in mean temperature of 2 °C over the Alps could be accompanied by large increases in extreme precipitation events (up to 30% increases in the southern part of the Alpine chain).

Detailed studies such as those reviewed here would be necessary elsewhere in the mountain regions of the world, primarily to characterize their current climatological characteristics, but also as a means of providing information useful to the

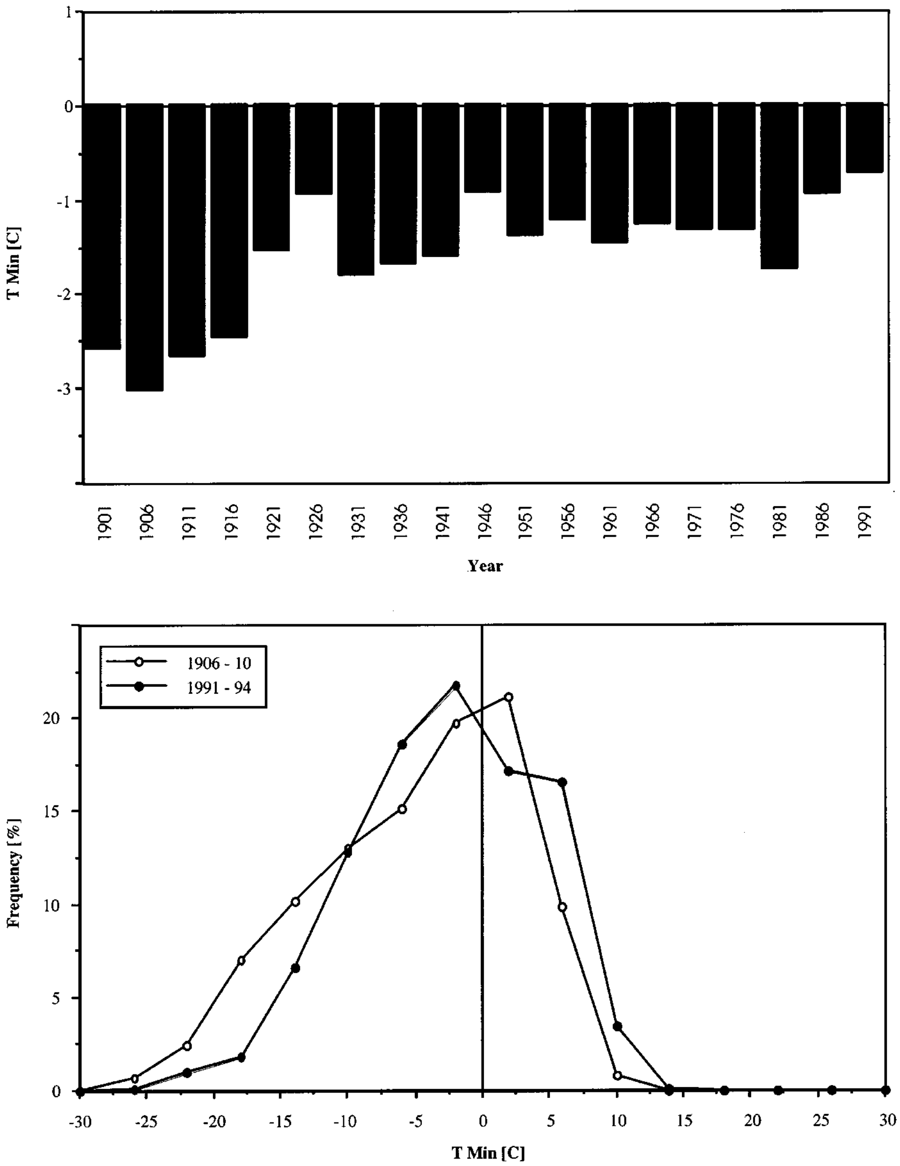

Figure 3. *Upper*: Five-year means of minimum temperature this century at Davos. *Lower*: Shift in the minimum temperature distribution profile at Davos as a function of the changes in the mean value; two profiles are illustrated, corresponding to the warmest (1991–1994) and coldest (1906–1910) periods of the century, respectively.

climate modeling community. Because of the expected refinements in the physical parameterizations of climate models in coming years, and the probable increase in the spatial resolution of GCMs, the use of appropriate data from high elevation sites will become of increasing importance for model initialization, verification, and intercomparison purposes. The necessity of accurate projections of climate change is paramount to assessing the likely impacts of climate change on mountain biodiversity, hydrology and cryosphere, and on the numerous economic activities which take place in these regions.

3. Climatic Variability: Observations and Model Simulations

The IPCC (1996) report documents much of the observed variations in surface temperature throughout the globe for about the past century, and provides a comparison of numerical model results of natural and forced climatic variability on a variety of time scales. Diaz and Graham (1996) have recently compared the behavior of observed freezing-level heights in the tropics to that obtained using the results of a numerical model simulation of the atmospheric response to observed sea surface temperatures (SST) in the period 1970–1988. The results show that the increase of tropical SST, which occurred in the mid-1970s, and has been documented in a number of studies (e.g., Trenberth, 1990; Trenberth and Hurrell, 1994; Miller et al., 1994), has led to increases in the height of the freezing-level surface in the tropics on the order of about 100 m, and consequently to warmer conditions in many high-elevation tropical ice caps, particularly in South America. This is consistent with the results of a number of studies of high-elevation tropical glaciers and ice caps (cf. Schubert, 1992, in Venezuela; Hastenrath and Kruss, 1992, in Kenya), which have shown a retreat of such features, particularly duuring the last few decades (Oerlemans, 1994; Thompson et al., 1995a,b).

Figure 4 shows the changes in freezing-level heights averaged over 10 radiosonde stations in South America. The figure also shows the changes in global tropical SST and in tropical precipitation in the Pacific Ocean for the past few decades. The graph illustrates two things. First, a significant upward change in the level of the 0 °C isotherm took place in the mid-1970s, coincident with changes in tropical SST and precipitation (see Diaz and Graham, 1996). Second, the change is also consistent with a global mid-tropospheric warming as documented from a global network of radiosonde stations (Angell, 1988; Oort and Liu, 1992; IPCC, 1996).

The results of Thompson et al. (1995b), Diaz and Graham (1996), and others, also suggest that high-elevation environments in the tropics may be particularly sensitive to changes in tropical SST, and that prolonged El Niño-like episodes, such as occurred recently from 1991 to early 1995 (Trenberth and Hoar, 1996), and decadal-scale changes, such as the mid-1970s Pacific Ocean episode are likely to impact the hydrologic and ecological balances of high-altitude zones throughout the globe, but perhaps most acutely in the tropics.

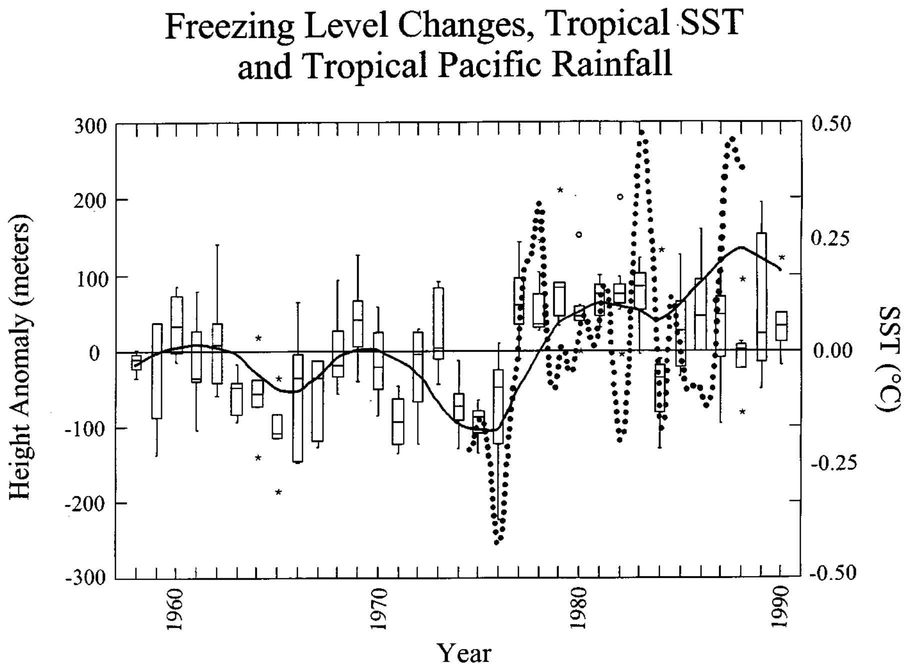

Figure 4. Distribution of annual freezing level height anomalies (geopotential height of the 0 °C surface) for 10 radiosonde stations in South America for the period 1958–1990. The smoothed line represents the low-pass filtered median tropical SST anomalies (seasonal values referenced to 1951–1992), based on a 5-degree gridded data set, and the solid dots correspond to the time coefficients (arbitary units, scaled to the right-hand-side) of the first principal component of tropical Pacific annual mean rainfall (after Diaz and Graham, 1996; principal component scores taken from Morrissey and Graham, 1996).

4. Holocene Paleoclimatic Changes in Mountain Regions

Because instrumental climate records from mountain areas are generally short, to obtain a perspective on climate variability before this century, we must rely on proxy records of past climate. In mountain regions such records are limited to lake and bog sediments, ice cores and glacier moraines, tree-rings, tree macrofossils and (for the last few centuries) documentary records (Table I). Each type of record has its own attributes, and limitations, so it is prudent to look for a convergence of evidence before relying too much on one particular reconstruction. Here, we provide a brief overview of the main approaches to paleoclimatic reconstruction in mountain areas, and summarize some of the principal conclusions such research has produced so far. Regrettably, most of the literature on mountain paleoclimates pertains to North America and Europe and this is reflected in our overview.

In the highest mountains, ice caps and glaciers are dominant features and in a few areas, ice cores have been recovered from mountain ice caps, providing a set of unique, high resolution records extending back for approximately 100 to over

Table I

Summary of paleoclimatic data types, the climate parameters they are a proxy for, and the spatial and temporal resolution typically associated with each of them. Also listed are some of the possible limitations associated with these proxy data sources

Source	Trees		Ice caps		Lakes and bogs		People
Data type	Rings	Macro-fossils	Ice cores	Glaciers	Sediments	Peat	Documents
Parameters	Width density isotopes (C, O)	Presence/ absence width density isotopes (C, O)	$\delta^{18}O$ = T? Accum = P (^{10}Be = R?) Air mass particles Gas content	Terminal position (volume)	Sediment character Geochem. biota pollen	Pollen	Multiple description
Resolution	Seasonal to annual	Irregular	Seasonal to decadal/ century	10^2–10^3	10^2–10^3 10 if varved	10^2–10^3	Daily $<$10 years
Time-frame: *Common* *Occasional* *Rare*	 500 1,500 10,000		 1,000 $>$5,000 $>$10,000	 1,000 to 10,000+	 to 10,000 yr BP		 500 1,000
Potential limitations	Frequency-limited, dependent on individual sample length and growth function; mostly multi-decadal to century time periods		Melt may destroy record; establishment of chronology problematical	Influence of topography and glacier dynamics. Extracting climate information difficult	Changes in upstream hydrology (glacier variation?) may confuse	Regional vs. local airflow may confuse	Verifying observational veracity; short-term, questionable low-frequency information

10,000 years. High elevation ice cores provide invaluable paleo-environmental information to supplement and expand upon that obtained from polar regions. To date, four high altitude sites have yielded ice cores to bedrock – Quelccaya and Huascarán in Peru, and Dunde and Guliya Ice Caps in western China. Where records extend back to the last glacial period (as in the Dunde and Huascarán ice cores) glacial stage ice is thin and close to the base making a detailed interpretation very difficult (Thompson et al., 1988a, 1989, 1990, 1995a, b). Nevertheless, even short sections of deep ice cores can yield important information. For example, in ice cores from the col of Huascarán, Peru (6,048 m) the lowest few meters contain ice from the last glacial maximum, with δ^{18}O $\sim$ 8% lower than Holocene levels, and a much higher dust content (Thompson et al., 1995b). The lower δ^{18}O suggests that tropical temperatures were significantly reduced in the LGM (by $\sim$ 8–12 °C), which supports arguments that changes in tropical SSTs were much greater than those indicated by the reconstructions of CLIMAP (1981) which have guided thinking on this matter for many years.

Because of the high accumulation rates on mountain ice caps, high elevation ice cores can provide a high resolution record of the recent past, with considerable detail on how climate has varied over the last 1,000–2,000 years, in particular (Thompson, 1991, 1992). The Quelccaya ice cores have been studied in most detail over this interval (Thompson et al., 1985, 1986; Thompson and Mosley-Thompson, 1987). Two cores extend back $\sim$1500 years (though only one can be reliably interpreted before $\sim$A.D. 1200). These reveal a fairly consistent seasonal cycle of microparticles, conductivity and δ^{18}O which (collectively) have been used to identify and date each annual layer. Dust levels increase in the dry season (June–September) when δ^{18}O values and conductivity levels are highest, providing a strong annual signal. A prominent conductivity peak in A.D. 1600 (associated with a major eruption of the Peruvian volcano Huaynaputina in February–March, 1600) provides an excellent chronostratigraphic check on the annual layer counts.

δ^{18}O over the last 1,000 years shows distinct variations in the Quelccaya core, with lowest values from A.D. 1530–1900. Accumulation was well above average for part of this time (1530–1700) but then fell to low levels. The overall period corresponds to the so-called ‘Little Ice Age’ observed in many other parts of the world. Accumulation was higher prior to this interval, especially from A.D. 600 to 1000. Archeological evidence shows that there was an expansion of highland cultural groups at that time. By contrast, during the subsequent dry episode in the mountains (A.D. 1040–1490) highland groups declined while cultural groups in coastal Peru and Ecuador expanded (Thompson et al., 1988b). This may reflect longer-term evidence for conditions which are common in El Niño years, when coastal areas are wet at the same time as the highlands of southern Peru are dry. Indeed, the Quelccaya record shows that El Niños are generally associated with low accumulation years, though there is no unique set of conditions observed in the ice core which permits unequivocal identification of an ENSO event (Thompson et al., 1984). Nevertheless, by incorporating ice core data with other types of proxy record it may be possible to constrain long-term reconstructions of ENSO events (e.g., Baumgartner et al., 1989; Michaelson and Thompson, 1992).

High altitude ice cores reflect significant increases in temperature over the last few decades, resulting in glaciers and ice caps disappearing altogether in some places (e.g., Schubert, 1992). This is quite different from polar regions where temperatures have declined in many regions during the same period. At Quelccaya, temperatures in the last 20 years have increased to the point that by the early 1990s melting had reached the Summit core site (5,670 m), obscuring the detailed δ^{18}O profile that was clearly visible in cores recovered in 1976 and 1983 (Thompson et al., 1993). In the entire 1,500 year record from Quelccaya, there is no comparable evidence for such melting at the Summit site. Similarly, at Huascarán, in northern Peru, δ^{18}O values increased markedly, from a 'Little Ice Age' minimum in the 17th–18th centuries, reaching a level for the last century which was higher than at any time in the last 3,000 years. Ice cores from Dunde Ice Cap, China, also show evidence of recent warming; δ^{18}O values are higher in the last 50 years than in any other 50 year period over the last 12,000 years (though the resolution of short-term changes in amplitude decreases with time). These records, plus evidence from other short ice cores from high altitudes (Hastenrath and Kruss, 1992) point to a dramatic climatic change in recent decades, prompting concern over the possible loss of these unique archives of paleo-environmental history (Thompson et al., 1993). The cause of the recent warming remains controversial.

Ice marginal positions are recorded by moraines, and trim-lines on valley walls. In ideal situations, it may be possible to identify a series of overlapping or nested moraines representing former glacier positions. However, more commonly the most recent advance of ice (the exact timing of which may have varied) has obliterated evidence of earlier advances because it was the most extensive for several millennia (and in some glaciers, the most extensive since the last ice age). These 'Little Ice Age' moraine systems are the latest in a series of glacier advances which began in the late Holocene, and which are collectively referred to as 'neoglaciations' (Grove, 1988; Matthews, 1991). Dating such advances is problematical, relying principally on radiocarbon dating of organic material buried by the advancing moraine, or by lichens growing on the moraine itself, once it has stabilized. Clearly, such evidence can only be episodic and does not provide the kind of high resolution, continuous data that is favored in paleoclimatic analysis. However, closely related records may be obtained from glacier-fed lakes which may register the growth of ice and the deterioration of mountain climates by a reduction in organic matter, and an increase in silt input to the lake-bottom sediments (Karlén, 1976, 1981; Nesje et al., 1991). In ideal circumstances, such records may be annually laminated, providing very high resolution insight into past climatic conditions (e.g., Leonard 1986). Pollen and other microfossils in lake sediments, or in high altitude bogs, can be interpreted in terms of former tree-line movements and hence provide a framework for other proxy records in the mountains (e.g., Burga 1993). Based on a composite view of such data, Karlén (1993) argues that glaciers (in the more continental parts of Scandinavia) advanced to positions comparable to those of the Little Ice Age around 3000, 2400, 2000 and 1200 years B.P. Many of these glaciers had completely disappeared in the early to mid-Holocene, only reforming within the last 3,000 years (Matthews, 1993; Nesje et al., 1994).

Beyond the realm of snow and ice, and alpine tundra, the tree-line defines an important climate-related ecotone. Although the tree-line itself varies in structure and composition from one mountain region to another, and is subject to many potentially limiting ecological constraints (Tranquillini, 1993) climate is the dominant control, at least away from the oceanic margin. Consequently evidence of past changes in tree-line position is generally interpreted in terms of variations in summer temperature. Radiocarbon-dated macrofossils (tree stumps, or wood fragments) from above the modern tree-line can thus provide dramatic testimony of warmer conditions in the past. This is well-illustrated in the northern Urals where now dead trees beyond the modern tree-line have been dendrochronologically dated to obtain information on the timing of past tree growth at high elevations (Shiyatov, 1993). This reveals that most of the trees were growing in the 10th–12th century A.D.; no trees were found to date from the late 18th and 19th centuries, indicating tree-line had retreated at that time. This evidence is strongly supported by tree-ring studies in nearby forests, where maximum ring widths were found at the time the forest advanced and minimum ring widths were characteristic of the 18th–19th centuries (Graybill and Shiyatov, 1992).

Sub-fossil wood from above the present tree limit has been found over wide areas of Scandinavia, and the mountains of the western U.S. (Kullman, 1989, 1993; Kvamme, 1993; Rochefort et al., 1994). In both areas, there is strong evidence that the upper tree limit was well above modern levels, especially before $\sim$5000 yr. B.P. In parts of the western U.S. and western Canada, trees were growing as much as 150 m above modern limits in the period from 8000 to 6000 yr B.P. (Rochefort et al., 1994) and similarly, in Scandinavia, trees were up to 300 m above modern limits in the early Holocene suggesting that summer temperatures were 1.5–2 °C above modern levels (Kvamme, 1993; Karlén, 1993). In both areas, it appears that climate deteriorated after 5000 yr B.P., leading to a decline in tree limits. This corresponds to both pollen records and the evidence from glacier moraines that temperatures became lower, especially after $\sim$3500 yr B.P., marking the onset of late Holocene neoglaciation. Minor oscillations of tree-line have taken place since then, culminating in the coldest episodes which we collectively term the ‘Little Ice Age’, from the 16th–19th century A.D. At that time, temperatures in the mountains of southern Sweden were $\sim$1 °C colder than in the mid 20th century (Kullman, 1989).

The overall picture from diverse paleoclimatic records in mountain areas is thus of early Holocene warmth, reaching an optimum around 6000 yr B.P., followed by a cooler late Holocene. The period after 5000 yr B.P. was punctuated by a few warm periods but there were also several especially cold episodes when glaciers advanced and tree-line declined. In terms of the current debate over anthropogenic versus ‘natural’ climate forcing, it is important to note that the instrumental record which we now use to characterize ‘global warming’ began at what was arguably the coldest period of the Holocene, in the mid-19th century. Clearly, pronounced climatic variations have been registered by proxy records in mountain areas long before any significant anthropogenic effects on global greenhouse gas concentrations, yet the causes of such variations remain obscure. Wigley and Kelly (1990) and Magny (1995) point out the correlation between Holocene glacier fluctuations and ^{14}C variations, a proxy which they consider to be indicative of solar irradiance variations. However the ^{14}C anomaly record is influenced by several factors, including changes in deepwater circulation of the ocean; it may be that the apparent links between ^{14}C and glacier fluctuations reflect subtle changes in the thermohaline circulation of the ocean, which clearly will influence the atmospheric circulation on a global scale. There are others who believe that the record of mountain glacier fluctuations (at least for the last 1,000 years) is closely linked to the level of volcanic aerosols in the atmosphere (e.g., Porter, 1986) and by inference, similar events earlier in the Holocene may also be indicative of higher volcanic dust loading of the atmosphere (Nesje and Johannessen, 1992). Whichever of these factors has been important, there is little doubt that on the very long timeframe, changes in radiation receipts, related to changes in the earth's orbit relative to the sun, have been the dominant factor controlling mountain climates over the last 20,000 years and it is probable that the observed record of higher tree-line in the early Holocene was directly attributable to higher radiation receipts and warmer summer temperatures at that time. The fairly rapid shift to lower tree-line, and the onset of neoglaciation soon after $\sim$5,000 B.P. in many areas seems to have been too abrupt to be simply due to declining radiation receipts, and some of the other factors mentioned may have contributed, individually or collectively in the cooling that set in then. Later cool episodes, culminating in the series of cold spells of the 'Little Ice Age' (16th–19th centuries A.D.) may be related to reduced solar irradiance and/or enhanced volcanic activity, though the possibility of a reduction in North Atlantic thermohaline circulation (affecting Europe especially) cannot be ruled out (Rind and Overpeck, 1993; Keigwin, 1996).

5. Conclusions

Climatic change at high elevation sites as observed in the paleo record and during this century is characterized by a high degree of complexity, associated with the orography itself, and a high degree of uncertainty because of the problems related to lack of observations and the difficulty in taking into account mountains in numerical climate models. However, in view of the unique environmental significance of mountains, particularly their biological and hydrological resources which govern numerous economic activities often well beyond the boundaries of the mountain areas themselves, it is essential that more effort be invested into understanding the climate characteristics of diverse mountain regions and the links between global climate forcing and regional climate response therein. There are clear indications from a number of high elevation climate records, that the amplitude of temperature changes this century is greater than the observed global or hemispheric change; furthermore, a number of lines of evidence suggest that the warming signal in the tropics during past few decades is amplified with height. It is likely, therefore that the impacts of future, accelerated climatic change will be proportionally more perceptible at high elevations. Small shifts in precipitation patterns in a specific mountain range, for example, could lead to widespread disruption of fresh water supply for agricultural, industrial and domestic use in regions far removed from the mountains but which are dependent on them for the resources they provide.

The IPCC Second Assessment Report (IPCC, 1996), in its chapter on the impacts of climate change on mountain regions (Beniston et al., 1996) has recommended that ‘future research needed to understand and predict effects of climatic change on mountain regions should represent balance and coordination between field studies (including paleoenvironmental data collection), monitoring, experimental studies, and modeling’. The areas of focus for future research suggested by the IPCC include, *inter alia*, the encouragement of specific regional field studies (transects, data acquisition, mapping, observations at high-elevations, etc.); paleo data to establish baselines and to evaluate the responses of ecosystems to natural climate variability, as well as to provide data for model verification; and monitoring programs to establish long-term baseline data, in particular in potentially sensitive regions (remote areas, high elevations). In the climate modeling domain, there is an urgent need to improve climate scenarios using various downscaling approaches; to improve understanding of how topographic and edaphic variability influence ecosystems and natural resources on the regional scale; and to improve the modeling of physical, biological and socio-economic systems.

It is by no means certain whether these recommendations will be followed, as they imply financial investments not always forthcoming in the present world economic climate. Among the unique historical archives of climate data for the world, those compiled at a number of mountain observatories, most notably in Europe, have the longest and most useful records. Unfortunately, a few of these unique observation series have been discontinued for budgetary reasons (e.g., Dessens and Bucher, 1995). It is nevertheless hoped that, in the context of debates, controversies, and uncertainties related to anthropogenic climate change, continuing support for the operation and maintenance of high-elevation observing sites will be sustained.

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