MOUNTAIN CLIMATE CHANGE AND CRYOSPHERIC RESPONSES: A REVIEW

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Introduction

Direct observations and proxy records indicate that historical and recent changes in climate in many mountain regions of the world are at least comparable with, and locally may be greater than, those observed in the adjacent lowlands. Actual and potential responses in cryospheric variables include: a rise in the snowline, a shorter duration of snow cover, changes in avalanche frequency and characteristics, glacier recession, break-out of ice-dammed lakes, warming of perenially -frozen ground and thawing of ground ice. Examples of these changes are presented from various mountain regions. The need for more and better observations for monitoring purposes is also addressed.

The changes – including the loss of ice core records of climate history as tropical glaciers and ice caps warm and melt water destroys the ice stratigraphy – are of scientific importance. There are also critical socio-economic implications. These include direct effects of the changes on water resources and hydropower generation, on slope stability and on hazards relating to avalanches and glacier lakes. Indirect effects include economic and social costs for winter tourism based on skiing and associated sports, and impacts on agricultural, industrial and consumptive use of water that is strongly influenced by the annual cycle associated with snow and ice melt runoff.

Evidence for changes in climate in mountain regions

Global mean annual temperature has risen by just over 0.6° C over the last century (Jones *et al.*, 1999), with accelerated warming in the last 10-15 years. The evidence for changes in climate in mountain areas is both direct and indirect. Observational records are available from the late nineteenth century at a number of mountain observatories, mostly in Europe (Barry, 1990). They indicate that mean temperatures have risen by amounts generally comparable with those observed in the lowlands during the twentieth century; however, there are some differences in the pattern of seasonal and diurnal changes. There are also differences in altitudinal gradients depending on the winter mean temperature. In Switzerland, from 1979-93, anomalies of minimum temperature increased (decreased) with altitude in mild (cold) winters (Beniston and Rebetez, 1995).

In the tropical Andes, mean annual temperature trends have been determined for 268 stations between 1° N and 23° S, for 1939-98 (Vuille and Bradley, 2000). They show an overall warming of about 0.1° C/decade, but the rate tripled to + 0.32-0.34° C/decade over the last 25 years. Trends vary with altitude, but there is generally reduced warming with elevation. This is especially apparent on the western (Pacific) slopes of the Andes.

There is a rather widespread tendency towards diurnal asymmetry, with the largest increases generally observed in nocturnal minimum temperatures, while daytime maxima have risen less or may even have decreased. Thus, the diurnal temperature range has diminished (Weber *et al.*, 1994). In a survey of available high elevation data, Diaz and Bradley (1997) present changes in zonally averaged temperatures for 1951-89 between 30°-70° N, versus elevation. Mean maximum temperatures increased slightly between 500 and 1500 m, with minor changes at higher elevations, while minimum temperatures rose by about 0.2° C/decade at elevations from 500 m to above 2500 m. In the Rocky Mountains, Pepin (2000) documents altitudinal differences in the changes in the Colorado Front Range since 1952, with overall cooling at 3750 m but warming between 2500 and 3100 m. This results in complex changes in lapse rates. Earlier, Brown *et al.* (1992) demonstrated that lapse rates between the high plains (1200-1500 m) and three stations at 3200 m in the Colorado Rocky Mountains had become shallower in the daytime, but steepened at night. Globally, the decrease in diurnal temperature range is attributed to increased cloud cover, locally augmented by changes in precipitation and soil moisture (Dai *et al.*, 1999).

An analysis of lapse rates in the Pennines of northern England indicates that atmospheric temperature and moisture level, cloudiness/solar radiation and wind speed determine lapse rates. Thus, lapse rate changes are complex and may result solely or partly from changes in the frequency of cyclonic/anticyclonic circulation regimes. A shallower/steeper lapse rate may be expected under warmer, moister atmospheric conditions or with increased solar radiation. Shallower lapse rates may also occur in northern England under southerly airflow. The amplitude of diurnal change in lapse rates intensifies under anticyclonic conditions and slack pressure gradients (Pepin *et al.*, 1999).

In some mountain regions, monitoring of ground temperatures has begun recently. In the northern Tien Shan, permafrost ground temperatures have risen by 0.2-0.3° C over the last 25 years (Gorbunov *et al.*, 2000). The depth of seasonal freezing has not changed significantly in the low mountains, but there has been a decrease in the depth between 1400 and 2700 m, while above 3000 m the depth of seasonal freezing is increasing. In the Swiss Alps, Haeberli (1994) estimated permafrost warming by about 1° C between 1880 and 1950, then stabilizing, before accelerated warming in the late 1980s to at least 1992. However, a ten-year borehole record analyzed by Vonder Mühll *et al.* (1998) indicates that warming until 1994 was largely compensated by rapid cooling in 1994-96.

In addition to instrumental measurements, there are also historical eyewitness accounts, documentary records (Pfister, 1985) and the testimony of landscape paintings of mountain glaciers in the Alps (Zumbuehl, 1988). The interpretation of such records requires careful content analysis and consistency checks. Nevertheless, they extend the records for the Alps back several centuries (Wanner *et al.*, 2000) and there are similar records for other parts of Europe (Pfister, 1995).

Proxy evidence of climatic change is available from changes in glacier size dated by lichenometry and carbon-14, from tree-ring series, and from ice cores, *inter alia*. Numerous accounts from various mountain regions exemplify these results (Luckman, 1997; Luckman and Villaba, 2001; Solomina, 1999; Kaser, 1999). These sources become even more important in mountain regions that lack direct records, or where these are of short duration, as in the Andes and other tropical regions (Barry and Seimon, 2000). On Mount Kilimanjaro, East Africa, for example, mapping from various surveys and photographs indicates substantial thinning and a reduction of 75 percent of the ice area between 1912 and 1989 (Hastenrath and Greischar, 1997). For a number of tropical upper-air meteorological sounding balloon stations, Diaz and Graham (1996) report a rise of 100-150 m in the altitude of the freezing level in the atmosphere over the inner tropics (10° N – 10° S) between 1970 and 1986, which is correlated with a warming in the sea surface over the eastern tropical Pacific. Thompson *et al.* (1993) show that on the Quelccaya Ice Cap, at 14° S in Peru, melt water penetration had obliterated the important climatic record provided by the famous ice core collected earlier from that location. It has also been suggested that changes in the biological evidence of anuran species in Costa Rica are linked with warming in the tropics and associated changes in humidity that could cause a rise in the elevation of cloud base, potentially disrupting the stability of the cloud forest zone (Pounds *et al.*, 1999).

Cryospheric responses

The effects of global warming on the cryosphere in mountain areas are most visibly manifested in the shrinkage of mountain glaciers and in reduced snow cover duration. However, the responses are by no means linear. For example, warmer winters imply higher atmospheric moisture content, and more snowfall is associated with an overall increase in precipitation. Records of glacier length and mass balance during the second half of the twentieth century show reductions in continental climatic regimes, but increases in maritime regimes, such as in Norway, southern Alaska and coastal areas of the Pacific Northwest in Canada and the United States. In the Tropics, the rise in freezing level noted above, as well as changes in atmospheric humidity and perhaps cloudiness, in some cases, has given rise to progressive reduction in mountain glaciers and ice caps over the last century. Particularly dramatic changes are evident in East Africa where there has been a linear decrease in ice area on Mount Kilimanjaro since the early 1900s (Hastenrath and Greischer, 1997). Essentially all the ice cover on East African summits will be lost within twenty years or so, unless there is a dramatic shift in climatic conditions.

In a more subtle example of changes in snow cover in Austria, Bohm (1986, p. 49) reported a reduction in May-September snow cover at Sonnblick (3106 m) from 82 days during 1910-25 to 53 days in 1955-70. The mean summer temperature was about 0.5°C higher in the second interval. However, the associated change in snow cover duration estimated from average gradients of snow cover duration and temperature lapse rate would only be about 10-11 days (Barry, 1990). Such non-linear responses may arise through local albedo-temperature feedback effects, but this still requires thorough investigation.

Large responses are also expected in terms of the annual hydrologic regime of rivers where the runoff comprises a significant proportion from snow cover melt and from ice wastage in heavily glacierized basins. Runoff models under global warming scenarios project a higher and earlier peak of spring runoff from snowmelt and reduced flow in summer (Rango and Martinec, 1998). For the upper Rhône, for example, Collins (1987) found discharge correlated with mean summer temperature: a 1° C cooling between 1941-50 and 1968-77 led to a 26% decrease in mean summer discharge. Warming trends will have the opposite effect, but a dominant component of runoff change in heavily glacierized basins is attributable to the reduction in ice area. Chen and Ohmura (1990) calculated an 11% decrease in runoff from a basin of the upper Rhône drainage with 66% ice cover between 1922-29 and 1968-72, compared with only 6% decrease in one with about 17% ice cover between 1910-19 and 1968-72. In the latter case, the Rhône at Porte du Scex, runoff changes also responded to changes in climate but a decrease in basin precipitation was offset by the effect of warmer summers increasing the ice melt.

Socio-economic consequences

Socio-economic effects of changes in mountain snow and ice characteristics will be both direct and indirect. Direct effects associated with a shorter snow season and shallower snow cover will include the reduction or loss of winter sports facilities, or the necessity for increased reliance on snowmaking capabilities, with attendant losses of income and costs of adaptation. For the Austrian Alps, losses will be exacerbated at lower elevations (Breiling and Charanza, 1999). Secondary effects resulting from these changes may include the loss of related service activities and income at mountain resorts. Summer tourism may also be affected as scenic mountain glaciers shrink and waste away. Maintaining tourist access to the terminus of the Upper Grindelwald glacier, in retreat since the mid-1980s, for example, has necessitated the construction of a wooden stairway.

The changes in snow melt runoff and its timing will have direct impacts on hydropower generation and impose requirements for alternative power sources. Power outages and loss of revenue by utility companies may be expected, depending upon the relative contribution of hydropower to total electricity generation. In adjacent lowland areas where spring runoff is a major source of water for irrigation and for stocking reservoirs, there may be even greater economic consequences. Changes in snow pack will also affect soil moisture levels in spring and summer with implications for soil biota, fire risk, and the productivity of mountain pastures and forests (Price and Barry, 1997).

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