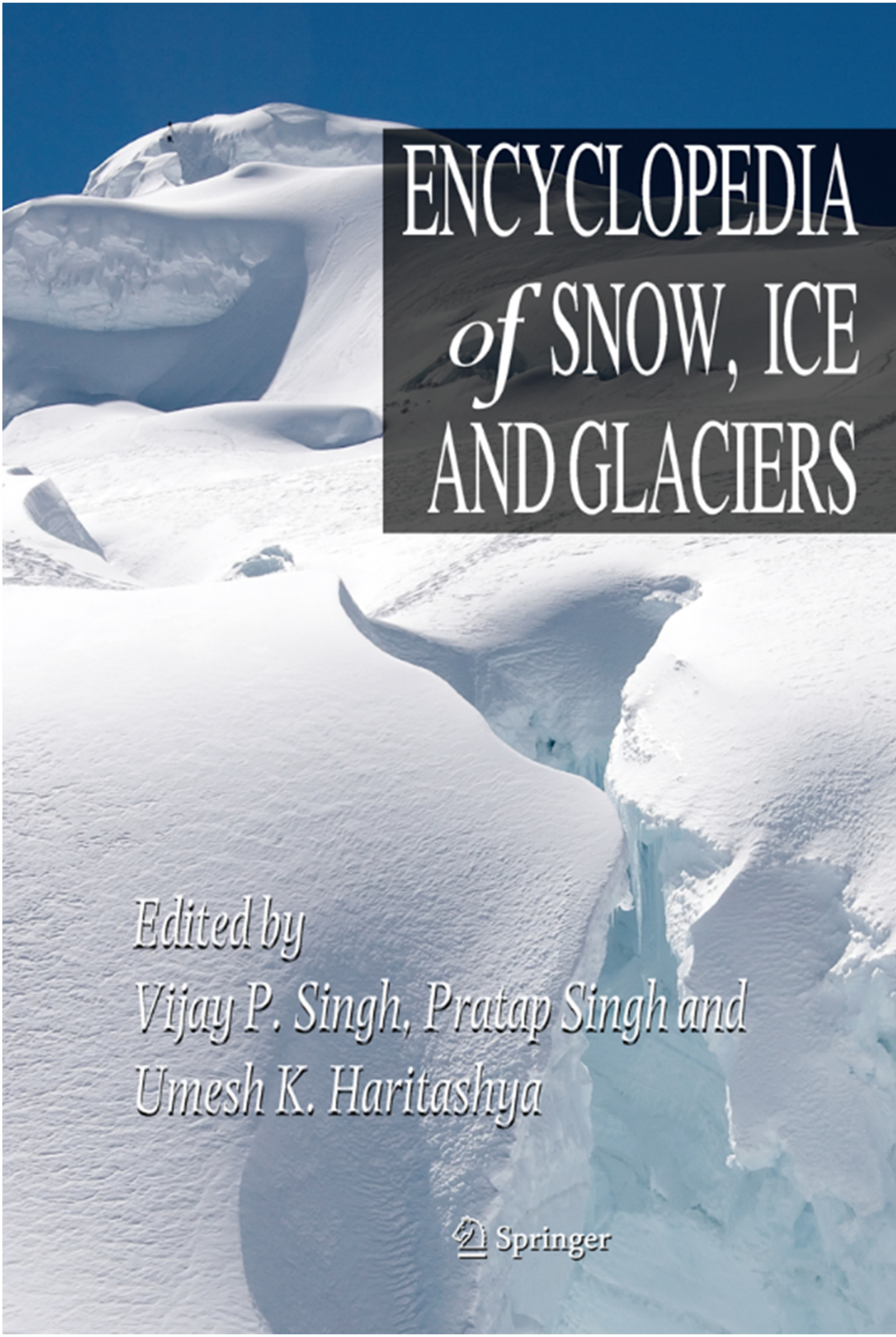

ENCYCLOPEDIA *of* SNOW, ICE AND GLACIERS

Edited by
Vijay P. Singh, Pratap Singh and Umesh K. Haritashya

Springer

ENCYCLOPEDIA *of* EARTH SCIENCES SERIES

Encyclopedia of Earth Sciences Series

ENCYCLOPEDIA OF SNOW, ICE AND GLACIERS

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CLIMATE VARIABILITY AND HIGH ALTITUDE TEMPERATURE AND PRECIPITATION

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Definition

High altitude temperature and precipitation variability: It is the inherent characteristic of precipitation and temperature to change over time. Variability is measured as the temperature or precipitation deviations (anomalies) over a given period of time from a climate statistic (long-term mean) averaged over a reference period.

Introduction

Mountains give rise to very distinct climates at their highest peaks where glaciers exist, and this mountain climate varies in ways that is quite different from nearby low elevations. Mountain-induced dynamic and thermodynamic processes modify synoptic weather systems and create regional-scale atmospheric circulation regimes that generate distinct wind systems, cloudiness, Precipitation patterns, etc., and lead to a very unique mountain climate (e.g., Barry, 2008). However, climate variability at high altitude (including temperature and precipitation variability) is not nearly as well understood as similar variations at lower elevations. The remoteness and difficulty in accessing many high elevation sites, combined with the complications of operating automated weather stations (AWS) at high elevations, make long-term measurements very challenging. Furthermore, the complex topography of high elevation sites often leads to very site-specific measurements that are not always representative of a larger regional mountain environment.

Climate observations at high elevation

While some high elevation observatories, in particular in the Alps, have maintained climate records for over 100 years, most mountain regions of the world are difficult to access and essentially devoid of any high-altitude observations. New advances in instrumentation, satellite telemetry, and power supply through solar panels have made high elevation measurements more feasible in these regions over the past decade. For example, new and unique measurements have become available from remote glacier sites in the tropical Andean Glaciers, the Himalayas, and Mt. Kilimanjaro, thanks to the installation of such AWS – locations, where previously no climatic information existed (e.g., Hardy et al., 1998, 2003; Georges and Kaser, 2002; Moore and Semple, 2004; Mölg et al., 2009). Figure 1 shows an example from an AWS installed and operated by the University of Massachusetts, Amherst on the summit of Quelccaya ice cap in Peru (14°S) at 5,670 m above sea level. Still, to be truly useful for climate research it is imperative that these AWS remain operational for several years and ideally decades to allow detection of trends and variability on interannual to decadal timescales (e.g., Bradley et al., 2004).

Climate Variability and High Altitude Temperature and Precipitation, Figure 1 Automated weather station (AWS) on the summit of Quelccaya ice cap, located at 14°S in the eastern Peruvian Andes (Cordillera Vilcanota) at 5,670 m above sea level. Photo courtesy of Douglas R. Hardy.

Characteristics of temperature and precipitation at high elevations

In the free atmosphere temperature decreases with height at a rate of about 6°C km^{-1} (Environmental Lapse Rate), although this rate varies by region, season, time of day, and by the type of air mass. Similarly the diurnal temperature range also decreases with elevation in the free atmosphere; an effect that can also be observed on mountain slopes and summits where mixing of slope air with the free atmosphere occurs (Barry, 2008). The comparison between near-surface observations at high elevations and measurements in the surrounding free atmosphere at the same elevation, however, is not straightforward (e.g., Pepin and Seidel, 2005) as temperature in the free atmosphere is generally colder than its near-surface counterpart due to both latent and sensible heating of the atmosphere above elevated surfaces. Therefore, temperature lapse rates on a mountain slope may bear a close resemblance to the free atmospheric lapse rate or may be almost independent (Barry, 2008).

Precipitation distribution and amount are also strongly affected by mountain barriers; but in many mountain regions the exact mechanisms and impacts on Precipitation are still poorly understood due to paucity of data and problems related to accurate measurements of snowfall totals, in particular at exposed, windy, high-elevation sites (e.g., Falvey and Garreaud, 2007). In general, high elevations sites are affected by mountain-induced Orographic Uplift or convective instability that lead to regionally enhanced Precipitation. In typical convective patterns, common on tropical mountains, Precipitation is usually highest near the cloud base (generally at or below 1,500 m) and decreases significantly at higher elevations. The zone of maximum Precipitation tends to occur at higher elevations in drier climates. In mid-latitudes where Precipitation is derived primarily from advective situations, at least during the winter season, forced large-scale ascent over a barrier can lead to enhanced Precipitation even at 3,000 m or above on the windward side, due to both higher intensity and longer duration of Precipitation events (Barry, 2008). On the leeward side, however, the remaining moisture that spills over the mountain crest is usually insufficient to induce significant condensation and Precipitation amounts tend to be much lower than on the windward side.

Climate variability and change at high elevation

In many mountain ranges of the world both Precipitation and temperature vary on interannual timescales in response to changes in the large-scale circulation, forced by major modes of ocean–atmosphere interactions. In the Alps, for example, winter precipitation is sensitive to the phase of the North Atlantic Oscillation, with decreased snowfall and higher temperatures during its positive phase (e.g., Beniston, 1997, 2006). Similarly Precipitation in parts of the Rocky Mountains, the Cascades, and the Alaskan coastal range (Alaskan Glaciers) are influenced by the Pacific Decadal Oscillation, while snowfall amounts in the mountains of East Africa and the Himalayas are sensitive to the phase of the Indian Ocean dipole and the El Niño–Southern Oscillation (ENSO) phenomenon (e.g., Vuille et al., 2005; Chan et al., 2008). Temperature and snowfall variations in the tropical Andean Glaciers are also primarily a reflection of ENSO variability (Vuille et al., 2000; Garreaud et al., 2003), while the southern Andean Glaciers are more strongly influenced by the state of the Antarctic Oscillation (Gillett et al., 2006).

Superimposed on these natural climate variations, caused by ocean–atmosphere interactions, are long-term trends in temperature and Precipitation that have become discernible at many high-elevation sites over the past decades. There is clear evidence from many mountain ranges that the temperature increase over the past 100 years has been significantly amplified at high elevations when compared with low elevations or the global average temperature (e.g., Beniston et al., 1997; Diaz and Bradley, 1997), and that the warming is more closely related to an increase in daily minimum temperature than a change in the daily maximum (Diaz and Bradley, 1997; Beniston, 2006; Giambelluca et al., 2008).

The differential temperature trends with altitude are particularly apparent in the Alps and on the Tibetan Plateau. In the Alps many locations have seen an increase in minimum temperature of 2°C or more during the twentieth century (Beniston, 2006). Liu and Chen (2000) reported a significant warming on the Tibetan Plateau since the 1950s (0.16°C per decade), and especially during winter (0.32°C per decade). They observed an amplified warming at higher elevations, which was later attributed primarily to a strong elevation dependence of trends in minimum temperature (Liu et al., 2009). However, this dependence does not seem to hold for temperature extremes (You et al., 2008). In the mountains of the western United States (Rocky Mountains) the strongest warming (0.5–0.6°C between 1950 and 2000) seems to have occurred below 2,000 m (Diaz, 2005), although strong summertime warming at high elevations has lead to a significant reduction of alpine tundra (Diaz and Eischeid, 2007). In East Africa the lack of an adequate observational network has so far precluded a definite assessment of temperature changes at high altitudes. While some suggest that temperature has also increased significantly at highest elevations of the East African Mountains (e.g., Taylor et al., 2006), this has been questioned by others (e.g., Mölg et al., 2006). In the tropical Andean Glaciers the observed warming is stronger at higher elevation only on the eastern slope, while on the western side the strongest warming is recorded close to sea level (Vuille and Bradley, 2000; Vuille et al., 2003). This differential response may be related to changes in cloud cover and the lack of a seasonal snow cover at high elevations, which precludes an amplified warming due to a snow-albedo feedback (e.g., Pepin and Lundquist, 2008). Nonetheless, temperatures at high elevations in the tropical Andean Glaciers have increased by about 0.68°C over the past 70 years (Vuille et al., 2008), consistent with the observed increase in the freezing level height (altitude at which air temperature is close to 0°C) of about 1.43 m $year^{-1}$ between 1948 and 2000 (Diaz et al., 2003). Much of the warming in the high elevation tropics and hence the increase in freezing levels can be traced back to warmer tropical sea surface temperatures SST (Diaz and Graham, 1996; Diaz et al., 2003; Vuille et al., 2003). In the southern Andean Glaciers of central Chile the freezing level has also increased by 122 m during winter and 200 m during summer between 1975 and 2001 (Carrasco et al., 2005), leading to a rise in the glacier equilibrium line altitude (Snow Line) (Carrasco et al., 2008).

Projections of future climate change under different Greenhouse Gas emission scenarios suggest that in many locations higher elevations will continue to experience the strongest warming (Global Warming and its Effect on Snow/Ice/Glaciers). Fyfe and Flato (1999) report that the strongest twenty-first century warming in the Rocky Mountains will occur at the highest elevations. Similarly model projections in the tropical Andean Glaciers suggest that both surface and free-tropospheric temperature changes will be largest at higher elevations where glaciers are located (Bradley et al., 2006; Vuille et al., 2008; Urrutia and Vuille, 2009). Simulations with regional climate models in the Alps also project a significant warming of 4–6°C by the end of the twenty-first century, when compared to the 1961–1990 average. In general winters will be warmer and more humid in the Alps, while summers will also be warmer, but drier than today (see Beniston, 2006 and references therein). For the Tibetan Plateau Liu et al. (2009) project increases between 2.9°C (below 500 m) and 3.9°C (above 5,000 m) by the end of the twenty-first century.

While there is strong evidence for warming in most mountain regions, the picture for changes in Precipitation is much more mixed. In the northwestern Alps winter precipitation has increased significantly during the twentieth century (up to 30% in the last 100 years) but decreased by the same amount in the southeast (Schmidli et al., 2005; Schär and Frei, 2005). Vuille et al. (2003) found a positive Precipitation trend in the tropical Andean Glaciers north of about 10°S and a negative trend further south, but in general the trends were weak and statistically not significant. Bhutiyani et al. (2009) reported a significant decline in summer monsoon Precipitation over the northwestern Himalayas during the past 140 years, but found no change in the amount of winter Precipitation. In general changes in timing or amount of Precipitation are much more ambiguous and difficult to detect and there is no clear evidence of significant changes in Precipitation patterns in most mountain regions. Nonetheless Precipitation characteristics at high elevation will change significantly over the next 100 years as the increase in temperature will lead to more Precipitation falling in the form of rain. For roughly every °C rise in temperature the snow/rain transition will rise by about 150 m (e.g., Beniston, 2003).

Summary

Temperature and Precipitation variability are still poorly understood at many high elevation sites due to the lack of an adequate long-term monitoring network, but studies on mountain-induced dynamic and thermodynamic processes have advanced our understanding of climate variability at high altitude. In many mountain ranges of the world large-scale ocean-atmosphere interactions are the main driver for observed variability in both Precipitation and temperature. Over the past 100 years long-term warming trends (Global Warming and its Effect on Snow/Ice/Glaciers) have been superimposed on this natural variability and become increasingly evident at most high altitude sites. In many mountain regions of the world high altitudes appear to experience a stronger warming than the surrounding lowlands. Projections of future climate change in the twenty-first century suggest continued warming and rising freezing levels, combined with altered Precipitation patterns in many high altitude locals.

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Cross-references

Alaskan Glaciers
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Global Warming and its Effect on Snow/Ice/Glaciers
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Rocky Mountains
Snow Line
Temperature Lapse Rates in Glacierized Basins
Tibetan Plateau

CLOUDBURST

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Cloudbursts have no strict meteorological definition. The term usually signifies a sudden, heavy fall of rain over a small area in a short period of time. Cloudburst represents cumulonimbus convection in conditions of marked moist thermodynamic instability and deep, rapid dynamic lifting by steep orography. The phenomenon occurs due to sudden upward drift of moisture-laden clouds as a tall vertical column termed "Cumulonimbus clouds." The ascending moisture-laden clouds become heavier and at certain point they produce violent rainstorm within a short interval. Orographic lifting of moist unstable air releases convective available potential energy necessary for a cloudburst (Das et al., 2006).