Snowline Altitude and Climate in the Peruvian Andes (5-17° S) at Present and during the Latest Pleistocene Glacial Maximum

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Abstract
The present snowline in the Peruvian Andes (5-17°S), rises from as low as 4.7±0.1 km on the eastern (windward) to more than 5.3±0.1 km on the western (leeward) side of the central Andes. The effect of temperature on snowline altitude is isolated from the effect of precipitation by subtracting the altitude of the mean annual 0°C isotherm from the altitude of the snowline. This difference, defined as the normalized snowline altitude, increases with decreasing precipitation.

The lowest late Pleistocene snowline rose from east to west and ranged in altitude from 3.2 to 4.9 (±0.1) km. Both the present and lowest late Pleistocene snowlines indicate that moisture at both times was derived principally from tropical easterly winds. An east-west precipitation gradient steeper than present is inferred for the eastern slopes of the central Andes from the steeper late Pleistocene snowline gradient. Mean annual temperatures were 10±1.9°C cooler that today at 3.52 km, as calculated from a late Pleistocene snowline as much as 1.4±0.2 km lower than today. Mean annual precipitation was 25 to 50% less than today along the eastern side, and more than 75% less on the western side of the central Andes. These estimates of lower temperature and decreased precipitation are more extreme than previous estimates. They imply that the amount of glacial-age cooling elsewhere, such as in western North America, may also have been underestimated by previous researchers because they did not adequately consider the effect of reduced ice-age precipitation on snowline lowering.

I. Introduction
This paper is a review of portions of the Ph. D. dissertation of the first author (Fox, 1993), summarized and edited by the second author, to illustrate some new techniques for estimating and interpreting present and former snowline altitude in mountain regions. Fox (1993) analyzed snowline altitudes in three regions of the Andes Cordillera of South America (Fig. 1). The single region chosen for discussion here is the portion of Peruvian Andes between latitude 5°S, and about 17°S, and westward of 69.5°W longitude (Fig. 1). The limits of the region were defined by the availability of good quality topographic maps at a scale of 1:100,000 with a contour interval of 50 m, compiled from stereographic aerial photographs. The methods illustrated here are applicable to any other mountainous region

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with maps of at least comparable quality, climatic records, and aerial photographs or other documentation of snowlines, both past and present. In the original study, Landsat Thematic Mapper images were extensively used to compile the extent of modern snowfields and to estimate the extent of snow cover during the latest Pleistocene glacial maximum. The
age of the latest glacial maximum in the Peruvian Andes is not known, but was probably about 18,000 to 20,000 years ago.

Many authors have discussed the probable effect of variable precipitation on estimates of snowline lowering and cooling during the last glacial maximum. In particular, if the glacial climate was generally drier than at present, the snowline on mountains would not have lowered as much as it would have if precipitation had remained the same and only a colder temperature had determined the amount of snowline lowering. Conversely, estimates of temperature decrease during the latest glacial maximum that are based on an assumption of no change in precipitation will be systematically too small. This creates special problems for interpreting temperature change in the tropics during the latest glacial maximum. Many analyses of tropical mountains predict snowline lowering on the order of 1,000 m during the latest glacial maximum; based on a moist adiabatic lapse rate of 6 °C/1,000 m, tropical mountain cooling of at least 6 °C is commonly inferred. If however, the climate was drier and the snowline was therefore lowered less, the cooling would have been even greater. The implications of this effect are discussed in the final section of this paper.

II. Present Snowline

Present perennial snow cover and the snowline altitude in the Peruvian Andes was determined from the snow cover shown on the topographic map series published at a scale of 1 : 100,000 by the Instituto Geografico Militar, Lima, Peru. The maps were produced from stereographic aerial photographs that were taken between 1955 and 1963. Snow cover was photogrammetrically transferred from the aerial photographs to the topographic maps during production and is shown by blue contour lines with a contour interval of 50 m.

Snow cover is unevenly distributed along the axis of the central Peruvian Andes and is notably concentrated on mountain peaks higher than 5 km elevation from 8.5 °S to 13.5 °S latitude, and along the eastern and western margins of the Peruvian Altiplano (Fig. 2). Clapperton (1991) gave a good general review of the tectonic and volcanic controls on the spatial distribution of high topography and glaciation in the Andes.

Polygons were digitized around the perimeter of all snow-covered areas shown on the topographic maps using the ESRI ARC/INFO geographic information system (GIS). Total snow cover in the Peruvian Andes, calculated by summing the area of the snow cover polygons, is approximately 3,700 km².

Point elevations were recorded where the perimeter of each snow-covered region crosses a topographic contour line to determine the local or orographic snowline altitude. Points were digitized at horizontal intervals of approximately 1 km to record local maxima and minima details of the orographic snowline. A total of 5,730 points were digitized, with elevations ranging from 4,050 m to 6,000 m.

The position of the snowline determined from the topographic maps is assumed to be close to its perennial position. Temporal differences in the snowline altitude are generally small in tropical mountains because snowline is primarily controlled by the altitude of the 0 °C isotherm, which has a smaller seasonal and interannual range in the tropics than in mountains at higher latitudes (Kuhn, 1981).

The regional snowline altitude was determined by digitally contouring the orographic snowline point observations (Fig. 2). Contours show the calculated mean altitude of the point measure-
ments within contiguous 20×20 km regions. The mean, rather than the maximum or minimum altitude, was selected because it removes the bias introduced by preferential melting or preservation of snow due solely to local slope aspect. Individual snowline observations generally vary from the mean by only 100 m or less, and are slightly higher on east-northeast than on west-southwest slopes because clouds, more common in the afternoon than in the morning, block insulation and reduce melting of snow on west-facing slopes (Hasenrath, 1967). Contours were interpolated between snow-covered areas using the altitude of snow-free peaks and Seltzer's (1987) glaciation threshold estimates as controls. The error in determining the altitude of the snowline is estimated to be ±100 m, based on the local variability of the orographic snowline altitude observed on the topographic maps.

Snowline contours generally parallel topographic contours, but rise from 4,700 m in the

Fig. 2 Extent of present snow and ice digitized from the Peruvian 1:100,000 scale topographic map series (Fox, 1993, Appendix A) Contours show the altitude of the present regional snowline in meters. Polygons located east of the available topographic map coverage represent the additional extent of snow cover observed on 1:250,000 scale Landsat MSS images.
northern and eastern Peruvian Andes to more than 5,300 m in the south and west (Fig. 2). Snowline gradients as steep as 11 m/km occur perpendicular to the axis of the mountain belt on the east side of the Altiplano at 14 °S latitude where regional topographic and precipitation gradients are also steep, but an east-west snowline gradient of 5 m/km is more representative of the northern region.

The snowline contours shown in Fig. 2 are similar in trend and magnitude to previous estimates of the present snowline altitude that were calculated from far fewer orographic snowline measurements. On the basis of about 30 observations, Nogami (1976) showed the present snowline increasing from 4,800 m in the northern Peruvian Andes to 6,000 m in the southwest. Seltzer (1987) contoured single values of the modern glaciation threshold that he determined for 57 of the 160 topographic maps utilized in this study and showed snowline contours increasing from 4,700 m in the north to 5,200 m in the southwest, including a local maximum of 5,200 m in the vicinity of the Quelccaya Ice Cap (14 °S, 70 °W). Glaciological investigations by Thompson (1980) showed that the annual snowline on the Quelccaya Ice Cap is at 5,250 m, which compares well with the value of 5,300 m shown in Fig. 2. A meridional transect along the eastern margin of the Peruvian Andes by Satoh (1979) showed the present snowline increasing from a minimum altitude of 4,700 m in the north to 4,900 m in the south. Meridional transects by Hastenrath (1967, 1971 b) showed comparable increases in the present snowline altitude from both north to south and east to west.

The largest differences between all of the previous estimates of snowline altitude and the snowline contours shown in Fig. 2 are on the western side of the Peruvian Altiplano where the previous reports estimated the snowline at elevations greater than 5,500 m. Landsat Multispectral Scanner images acquired between October 1972 and May 1982 and Landsat TM images acquired between July 1984 and September 1986 support the lower snowline altitudes shown in Fig. 2.

III. Climatic Controls on the Present Snowline Altitude

Many previous studies have sought to identify and understand the factors controlling the present snowline altitude so that paleoclimatic inferences can be made from measurements of lower snowlines in the past (Charlesworth, 1957; Østrem, 1966, 1972; Bradley, 1975; Chinn, 1975; Miller et al., 1975; Porter, 1975, 1977; Wright, 1983; Seltzer, 1987; Broecker and Denton, 1989; Fox, 1991). Temperature and precipitation have been identified for a long time as the primary climatic controls on snowline altitude. Topography influences the snowline in alpine regions by its indirect control on temperature as a function of altitude and on precipitation as a function of the orographic lifting and blocking of moist air masses. Snowlines are influenced by other climatic variables, including cloudiness (Porter, 1975; Wright, 1983; Jordan, 1991), shortwave irradiance and sublimation (Johnson, 1976; Paterson, 1981), the proportion of rain to snow (Miller et al., 1975), and the effect of allochthonous snow accumulation due to wind drifting (Charlesworth, 1957). These variables are generally unquantified in the Peruvian Andes, as in many other remote alpine regions of the world, but their combined effects may be only minor and are probably related to the principal controls discussed below.

1) Temperature

Air temperature decreases non-linearly with
increasing altitude in the Peruvian Andes. Stone and Carlson (1979) found that temperature lapse rates in the lower tropical troposphere are within 20% of the non-linear, moist adiabatic lapse rate at altitudes above the planetary boundary layer (~900 mb). Therefore, a moist adiabatic temperature lapse rate describes the observed decrease in temperature with increasing altitude in the tropical Andes better than an assumed dry adiabatic lapse rate (linear) of 6.5 °C/km. Vertical transfer of heat in the tropics is dominated by moist convection in all seasons and there is little seasonal change in mean daily temperature. Since the moist adiabatic temperature lapse rate is dependent only upon land–surface temperature, it has very little seasonal change in the tropics.

Figure 3 shows the relationships between the elevation and mean summer, winter, and annual temperature for climate stations in the Peruvian Andes (Fox, 1993, Appendix B). Although an assumed or calculated linear temperature lapse rate close to 6.5 °C/km is often used to estimate the decrease in temperature with increasing altitude, such a linear fit to the data would overestimate temperatures both at sea level and at the elevation of the 0 °C isotherm. Second-order polynomial regression lines fit the temperature observations better than linear regression lines. The resulting equations, when differentiated, closely approximate the non-linear, moist adiabatic temperature lapse rates which predominate at tropical latitudes (Rennick, 1977). They are: \( y = 24.60 - 2.04 x - 0.54 x^2 \) (R²=0.94) Summer (Dec-Feb), \( y = 24.29 - 2.16 x - 0.65 x^2 \) (R²=0.94) Winter (Jun-Aug), \( y = 24.51 - 1.93 x - 0.62 x^2 \) (R²=0.94) Annual.

The annual and seasonal temperature lapse rates in northern (5–12 °S) and southern Peru (12–18 °S) were calculated separately and do not differ significantly from the equations shown above. Others have also noted that the moist adiabatic temperature lapse rate does not vary significantly with latitude in Peru (Has-
tenrath, 1971a; Johnson, 1976) These observations all confirm interpretations of radiosonde data by Hastenrath (1967) and Satoh (1979), and show that there is no latitudinal variation in the altitude of the annual 0 °C isotherm. However, the moist adiabatic temperature lapse rate does vary slightly throughout the year so that the mean altitude of the 0 °C isotherm ranges between 4,675 m in winter and 5,120 m in summer (Fig. 3). The altitude of the annual 0 °C isotherm is 4,920 m, and is similar to previous estimates of 4,950 m (Johnson, 1976) and 5,078 m (Seltzer, 1987) that were based on fewer climate station records, and estimates of 4,800 to 5,000 m (Hastenrath, 1967) and 4,783 m (Satoh, 1979) that were based on radiosonde data. The winter temperature lapse rate is steeper than the summer rate because the moist adiabat is inversely related to land-surface temperature (Stone and Carlson, 1979; Eagleman, 1985). As a result, higher elevations experience a larger annual range of temperature (Fig. 3). The altitude of the annual 0 °C isotherm interpolated from available climate station and radiosonde temperature data is believed to be within 100 m of its true position.

Despite an annual range of mean daily temperatures of less than 5 °C everywhere in Peru, the diurnal temperature range at high altitudes can exceed 20 °C and still maintain an average daily temperature at or below freezing. Although melting may occur daily at the snowline, winter is the ablation season in the Peruvian Andes because virtually no snow accumulates at this time, and the mass balance of snow and ice is negative.

Previous studies have identified the altitude of the 0 °C isotherm as the primary control on the altitude of the snowline in the tropical Andes (Hastenrath, 1967; Satoh, 1979; Kuhn, 1981, 1989; Seltzer, 1987). The lowest altitude of the snowline in the Peruvian Andes (about 4,700 m) is coincident with the altitude of the ablation season (winter) freezing level in the humid northern and eastern Peruvian Andes. The snowline rises westward from this level toward drier regions. In fact, the modern snowline is located at or above the ablation season freezing level everywhere in the tropics (Hastenrath, 1985; Allison and Peterson, 1989; Young and Hastenrath, 1991).

2) Precipitation

Moisture is transported to the Peruvian Andes primarily by humid easterly winds from Atlantic Ocean and Amazon Basin sources. As a result, the eastern side of the cordillera receives more annual precipitation than the western side (Fig. 4). More than 1,000 mm/yr of rain falls everywhere along the eastern margin of the Andes below an elevation of about 3,000 m, decreasing westward to less than 100 mm/yr along the Pacific coast. Local precipitation reaches maxima of 4,000 mm/yr and 6,400 mm/yr at about 8°S and 13°S respectively. Precipitation is produced and concentrated along the eastern side of the Andes as moist air is orographically lifted over the mountains, with consequent adiabatic cooling and condensation. Local slope aspect is also important, with windward slopes typically receiving more precipitation than leeward slopes (Johnson, 1976). High regional topographic gradients are responsible for high regional precipitation gradients. The east-to-west precipitation gradient is steep along the eastern side of the Peruvian Andes and has a horizontal gradient of about −15 mm/km on average and a maximum gradient of −145 mm/km at 14°S latitude. In contrast, the horizontal precipitation gradient is only −3 mm/km on the central Peruvian Altiplano to the southwest of the maximum precipitation gradient.
Because the annual 0 °C isotherm varies with latitude, a direct comparison of the relationship between precipitation and snowline altitude at different latitudes in the Andes is not possible. The snowline must first be normalized to the altitude of the annual 0 °C isotherm to remove the temperature influence and isolate the influence of precipitation. The term "normalized snowline altitude" has been introduced (Fox, 1993, p. 9) to describe the elevation of the snowline above or below the annual 0 °C isotherm, and is primarily a function of precipitation. Quantities similar to the normalized snowline altitude include a measure of the degree of continentality (Kuhn, 1981) and the vertical height range of the periglacial zone (Humlum, 1988).

The normalized snowline altitude has a systematic relationship with mean annual precipitation for 21 snow-covered regions in the tropical Andes between 5° and 20°S latitude (Fig. 5). The driest portion of the northern Chilean and Bolivian Andes, as well as the Peruvian Andes, are included in Fig. 5 to investigate the influence of the present range of precipitation extremes in the tropical Andes on

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Fig. 4 Mean annual precipitation (mm) digitized from Hoffman (1975). Squares show the location of all known climate stations. Shaded region is the land surface above 3 km.
the present normalized snowline altitude, and to identify possible modern analogs for paleoclimate inferred from the normalized snowline altitude during the late Pleistocene glacial maximum. The normalized snowline altitude in the Peruvian Andes was calculated by subtracting the altitude of the annual 0 °C isotherm (4,920 m) from the snowline altitude (Fig. 2) and in Bolivia by subtracting the altitude of the annual 0 °C isotherm (4,800 m: Satoh, 1979) from the observations of the snowline altitude compiled by Nogami (1976). The estimated error in the normalized snowline altitude is ±200 m because it is calculated from the altitudes of the snowline and the annual 0 °C isotherm, both of which have errors of ±100 m.

The mathematical relationship between the present normalized snowline altitude and mean annual precipitation is non-linear (Fig. 5) and is best quantified with a logarithmic equation of the form, \( N = b - a \log P \), where the normalized snowline altitude (N) decreases with increasing mean annual precipitation (P). Similar non-linear relationships between snow accumulation and glacier equilibrium line altitudes (ELA’s) have been reported by some workers (Liestøl, 1967; Kasser, 1973, 1977) and linear approximations by others (Porter, 1977). The logarithmic relationship between the present normalized snowline altitude and present mean annual precipitation provides the tool for using a past normalized snowline altitude to infer paleoprecipitation.

IV. Lowest Late Pleistocene Snowline

The few available radiocarbon dates of glacial deposits place the last glacial maximum in the Peruvian Andes between 24 and 12 ka, with
an earlier maximum before 43 ka (Seltzer, 1990). In addition, there is indirect evidence for a penultimate glaciation in Peru assumed to date between 130 and 170 ka (Clapperton, 1983; Wright, 1983). Minor late-glacial advances occurred between 12 and 10 ka, but there is little evidence for Holocene glacier activity except for the minor fluctuations associated with the Neoglacial interval of the last 5,000 years. Despite the lack of detailed chronologic control, the temperature and precipitation conditions associated with the most extensive late Pleistocene glaciation, referred to here as the latest Pleistocene glacial maximum, can be inferred from the lowest cirques even though their exact age is unknown.

Cirques identified in this study clearly range in age, commonly forming chains up a single valley where the highest ones may have been occupied by ice as recently as the end of the last century (Seltzer, 1987). The question of cirque age is not a critical factor in this study because the primary objective is to estimate the maximum change in climate recorded by the lowest cirques in each region, which are inferred to be contemporaneous. Although the absolute age of the lowest cirques is unknown, Satoh (1979) made an argument for the contemporaneity of the lowest cirques along a single mountain range and even on neighboring ranges because they occur at a remarkably uniform altitude (±100 m). The contour line marking the position of the abrupt change in stream gradient, generally also the contour line immediately below the tarn, defines the outer limit of the cirque and was used to determine the altitude of the cirque floor. An error of ±100 m is assigned to the regional snowline altitude based on the local variability of cirque floor altitudes.

A total of 11,108 point observations of cirque locations and altitudes were digitized and compiled from the topographic map series at a scale of 1:100,000 (Fig. 6). Cirque floor altitudes range in elevation from 3,200 m to 4,850 m, with basins less confidently identified as cirques as low as 3,150 m between 6° and 6.5° south latitude. These observations confirm an earlier report of cirque floor altitudes lower than 3,300 m in the northern Peruvian Andes (Seltzer, 1987). The lowest cirques, like the lowest present snowline, are located in the northern and eastern Peruvian Andes.

The altitude of the lowest late Pleistocene snowline was reconstructed by contouring the lowest cirque floor altitudes recorded in each 20 × 20 km region (Fig. 6). The lowest cirques were used because they formed under glacial climatic conditions most different from the present interglacial climate. The snowline altitude was calculated at a spatial resolution of 20 km in order to take advantage of the large number of cirque floor observations, although the snowline calculated at spatial resolutions lower than 20 km is not significantly different than that shown in Fig. 6.

Late Pleistocene glacial maximum snowline contours generally parallel topographic contours and present snowline contours, but rise from east to west with a horizontal gradient steeper than at present. The horizontal snowline gradient reached a local maximum of 46 m/km a 10° S during the late Pleistocene glacial maximum (Fig. 6), but a horizontal gradient of 20 m/km was more representative along the eastern side of the Peruvian Andes where the present horizontal snowline gradient is less than 11 m/km.

The altitude of the lowest late Pleistocene snowline presented in Fig. 6 supports earlier descriptions of the late Pleistocene snowline in Peru. Hastenrath (1967, 1971b) showed the
Pleistocene snowline rising from an altitude of about 3,800 m in the north to 5,000 m in the south along a snowline transect down the western side of the Peruvian Andes. Hastenrath (1971 b) also showed the late Pleistocene glacial maximum snowline rising from about 4,200 m to 4,500 m along two northeast-southwest transects across the Peruvian cordillera. Satoh (1979) showed the Pleistocene snowline increasing in altitude from a minimum of about 3,400 m in the northern Peruvian Andes to about 3,500 m in the south along the eastern side of the cordillera. His snowline is higher along the western side of the Peruvian cordillera,
reaching a maximum altitude of 4,800 m in the southwest. Snowline contours maps by Seltzer (1987) and Nogami (1976) are also supported by the much greater number of points plotted in Fig. 6. Both show a rise in the Pleistocene snowline from about 3,200 m in the north to about 4,800 m in the southwest.

V. Snowline Difference and Paleoclimatic Implications

The altitudinal difference between the present and a former snowline, sometimes referred to as snowline depression, is a standard basis for estimating the glacial-age climate in mountainous regions (Flint, 1971). The lower altitude of the late Pleistocene glacial maximum snowline compared with the present snowline clearly indicates that temperatures were lower in the Peruvian Andes. The lack of a precipitation lapse rate in most mountainous regions, however, has dissuaded many workers from making a quantitative estimate of paleoprecipitation from snowline difference. As a result, the entire observed snowline difference has too frequently been ascribed to paleotemperature without considering the effects of paleoprecipitation. The present study eliminates these traditional problems by avoiding the use of an assumed temperature lapse rate and by using the normalized snowline altitude instead of the absolute snowline altitude. Use of the normalized snowline altitude isolates the precipitation effect on the altitude of the snowline from that of temperature.

The difference in altitude between the present (Fig. 2) and the lowest late Pleistocene (Fig. 6) snowlines in the Peruvian Andes is variable (Fig. 7). A maximum snowline difference of 1,400±200 m is calculated for the eastern side of the Peruvian Andes, rapidly decreasing to a minimum of 600 m in the western cordillera and on most of the Peruvian Altiplano. The error of ±200 m assigned to the difference in snowline altitude is based on the combined estimated ±100 m errors in determining the latitudes of both the present and lowest late Pleistocene snowlines.

As is the case everywhere in the tropics today, the lowest altitude of the present snowline in the Peruvian Andes is close to the annual 0 °C isotherm, but never below the winter 0 °C isotherm (4,700 m), even where precipitation is high. Because the temperature lapse rate during the late Pleistocene glacial maximum could not have been significantly different from that of today in the tropics (Rind and Peteet, 1985; Betts and Ridgway, 1992), it can be assumed that the snowline was not below the winter 0 °C isotherm. Therefore, the winter 0 °C isotherm must have been lower than today by an amount equal to or greater than the
maximum snowline difference of 1,400 m.

An altitude of 3,520 m can be inferred for the annual 0 °C isotherm during the late Pleistocene glacial maximum from an annual 0 °C isotherm 1,400 m lower than today (Fig. 3). A mean annual temperature 10 °C cooler than today in the Peruvian Andes at an elevation of 3,520 m can be calculated from the difference between the present (10 °C) and glacial-age (0 °C) mean annual temperatures at that altitude. This method of estimating paleotemperature is preferred over the traditional method of multiplying snowline difference by an assumed temperature lapse rate because it does not presume knowledge of the temperature lapse rate during the late Pleistocene glacial maximum. The temperature lapse rate must be known or assumed, however, to calculate paleotemperature at altitudes above or below 3,520 m. An error of ±1.9 °C is calculated for the estimate of paleotemperature from a total uncertainty of ±300 m in determining the altitude of the annual 0 °C isotherm during the late Pleistocene glacial maximum, including uncertainties both in the snowline depression (±200 m) and in the present altitude of the annual 0 °C isotherm (±100 m).

The normalized snowline altitude during the late Pleistocene glacial maximum in the Peruvian Andes can be used to make an estimate of paleoprecipitation. The present normalized snowline altitude is controlled only by precipitation because the direct effect of temperature (annual 0 °C isotherm) on the snowline altitude

Fig. 8 Normalized snowline altitude (m) during the late Pleistocene glacial maximum
Values in parentheses are not associated with a contour line.
has been removed (Fig. 5). Assuming that the relationship between the normalized snowline altitude and mean annual precipitation was the same during the late Pleistocene glacial maximum as it is now, an estimate of paleoprecipitation can be made from the normalized snowline altitude. The normalized snowline altitude shown in Fig. 8 was calculated by subtracting the inferred maximum altitude of the glacial-age annual 0 °C isotherm (3,520 m) from the snowline altitude shown in Fig. 6. The normalized snowline altitude was lowest along the eastern side of the Peruvian Andes and rose to the west reaching a maximum of 1,200 m on the western margin of the Peruvian Altiplano.

An estimate of mean annual precipitation during the late Pleistocene glacial maximum (Fig. 9) can be calculated from the normalized snowline altitude (Fig. 8) and the inverted logarithmic relationship between the present normalized snowline altitude and mean annual precipitation (Fig. 5). Given the ±200 m uncertainty in determining the normalized snowline altitude, an error of ±25 % is assigned to estimates of paleoprecipitation. Mean annual precipitation ranged from more than 1,000±250 mm along the eastern side of the northern and central Peruvian Andes to less than 200±50 mm along the western side of the cordillera. The inferred high precipitation along the eastern margin of the Peruvian Andes demonstrates that the primary moisture source for late Pleistocene glaciers was from the east as it is today. The east-to-west precipitation gradient, however, was steeper during the late Pleistocene glacial maximum than at present, ranging from about −15 mm/km to −50 mm/km compared with a modern representative horizontal precipitation gradient of about −16 mm/km.

The percent decrease in mean annual precipi-
ication from present amounts during the late Pleistocene glacial maximum is shown in Fig. 10. Precipitation may not have been significantly less than today along the eastern side of the northern and central Peruvian Andes during the late Pleistocene glacial maximum, in contrast to a much larger precipitation difference to the south and west where it reached a maximum reduction of about 80% (or 20% of the present mean annual precipitation) on the Peruvian Altiplano. If the late Pleistocene glacial maximum cooling was actually greater than 10 °C at 3,520 m, then precipitation amounts (Fig. 9) would have been less and the percent difference in mean annual precipitation (Fig. 10) would have been greater.

VI. Comparison with Previous Paleoclimatic Estimates

A number of independent estimates of ice-age climates in the Peruvian Andes and elsewhere during the late Pleistocene can be compared with the results of this study. Recent literature reviews of pollen, lake level, and glacial evidence by Markgraf (1989) and Seltzer (1990) indicate that, in general, glacial climates were colder and drier in the Peruvian and neighboring tropical Andes during the late Pleistocene than at present. Simulations of the climate during the last glacial maximum by a number of general circulation models support this interpretation (Manabe and Hahn, 1977; Kutzbach and Guetter, 1986). However, the magnitudes of the temperature and precipitation reductions presented here are much greater than earlier estimates (Table 1). Early estimates of temperature difference by Hastenrath (1967, 1971a) and Satoh (1979) from snowline difference were only qualitative. Wright (1983) inferred a mean annual temperature 2 °C lower than today during the late Pleistocene from an assumed modern temperature lapse rate of 7 °C/km and a conservative estimate of snowline difference (300 m) in the cordillera west of Laguna Junin at about 11 °S. Seltzer (1987) examined the present relationship be-

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Table 1 Summary of climatic changes Inferred for the Peruvian Andes during late Pleistocene glaciations
tween snowline altitude and climate to estimate a temperature reduction of 6.5 ° to 7.0 °C near Nevado Huaytapallana at about 12 °S in the central Peruvian Andes. The maximum altitude of the annual 0 °C isotherm (3,520 m) inferred in this study from the altitude of the snowline during the late Pleistocene glacial maximum supports previous estimates in the tropical Andes. A lowering of the annual 0 °C isotherm from its present position down to about 3,500 m in Colombia was inferred from Pollen data by van der Hammen (1974) and by La Fontaine (1988) in the tropical Andes using the NCAR-CCM simulation of the last glacial maximum (Kutzbach and Guetter, 1986).

Previous estimates of paleoprecipitation from snowline difference in the Peruvian Andes vary significantly (Table 1). Hastenrath (1971a) and Wright (1983) reasoned that a larger snowline difference along the east side of the Peruvian Andes than to the west would have resulted from a southward shift of the equatorial zone of high precipitation associated with the ITCZ. The contemporaneous shift of the mid-latitude westerly wind regime northward during glacial intervals of the Pleistocene (Hastenrath, 1971a) would have effectively narrowed the subtropical arid zone from its present width. However, Satoh (1979) and Seltzer (1987) argued from snowline observations that conditions in the Peruvian Andes were drier than today during late Pleistocene glaciations and that snowline difference resulted primarily from a decrease in temperature. General circulation model simulations support their interpretation and do not show a southward shift in the position of the ITCZ during the last glacial maximum (La Fontaine, 1988). The fact that normalized snowline altitudes were higher in the Peruvian Andes during the late Pleistocene glacial maximum (−100 to 1,200 m) than at present (−100 to 400 m) supports an interpretation of more arid conditions (Figs. 5 and 8).

The climate of the last glacial maximum has been simulated by several groups using general circulation models (Table 1) and ice-age boundary conditions reconstructed by CLIMAP project members (1976, 1981). Early models by Gates (1976) and Manabe and Hahn (1977) used CLIMAP project members (1976) boundary conditions for July at 18 ka to estimate a reduction of mean July temperature by about 5 °C and little or no change in mean July cloudiness and precipitation in the central Peruvian Andes. The CLIMAP project members (1981) reassessment of the sea surface temperature data for 18 ka resulted in warmer temperatures than in the 1976 reconstruction. Subsequent models by Manabe and Broccoli (1985), Kutzbach and Guetter (1986), and Rind (1987), using the warmer CLIMAP project members (1981) sea surface temperatures, estimated less cooling (0 ° to 4 °C) than earlier model results, and seasonal decreases in precipitation of less than 4 mm/day (<38 %) from present conditions in the Peruvian Andes (Table 1). The decreases in temperature and precipitation simulated by general circulation models for the last glacial maximum are, therefore, much less than the decreases inferred from snowline difference.

Small changes in the climate of the last glacial maximum simulated by general circulation models cannot be reconciled easily with the much larger decreases in temperature and precipitation inferred in this study from snowline difference at high altitudes. One way to reconcile a 10 °C cooling at high altitudes (>3 km) with a 2 ° to 4 °C cooling at sea level is to invoke an unrealistic steepening of the ice-age temperature lapse rate. Rind and Peteet (1985)
stated, however, that a large divergence of the last glacial maximum temperature lapse rate from the moist adiabatic value is unlikely because moist convection is, and probably was, the dominant process of vertical heat transport in the troposphere at low latitudes. However, general circulation models do show a slightly steeper temperature lapse rate, greater cooling at higher altitudes than at lower altitudes, and an increased annual temperature range at 18 ka (Rind and Peteet, 1985; La Fontaine, 1988).

The discrepancies between inferred cooling and the effects of ice-age precipitation decrease are not restricted to the South American Andes. Porter et al. (1983, p. 78, 82) noted that late Wisconsin cooling in the Brooks Range and the Alaska Range has been underestimated because of decreased precipitation. Similar conclusions were reached for the western cordillera of the United States (Porter, et al., 1983, p. 103, 107). General application of the concept of a normalized snowline altitude, as presented in this paper, is likely to increase most previous estimates of mountain climatic cooling during the latest glacial maximum. The effect will be to increase the contradictions between CLIMAP project members (1981) estimates of sea-surface cooling and most estimates of cooling in mountain regions. Especially in the Andes, the cooling estimated in this paper (Table 1) is much greater than that inferred for the coast of Peru by the CLIMAP study. It appears that climatologists must undertake a major review of the CLIMAP report, which has served as such an important reference for almost two decades.

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