CHAPTER 2

MODERN AND LAST GLACIAL MAXIMUM SNOWLINES IN THE PERUVIAN-BOLIVIAN ANDES

Abstract

When glaciers were at their late Pleistocene maximum, snowlines in the central Andes of Peru, Bolivia and Chile were 500-1200 m lower than at present. Then and today, snowlines rise in elevation from northeast to southwest paralleling a precipitation decrease, indicating that regional atmospheric circulation patterns have not changed in this respect between the late Pleistocene to the present. Dates from several sites imply that the timing of the late Pleistocene glacial maximum in the tropical Andes occurred prior to 20,000 ^{14}C yr BP, but lack of maximum limiting ages adds considerable uncertainty to the timing.

Snowline modeling, based on the work of Kuhn (1989), demonstrates that snowlines in the eastern and western cordilleras of the Andes respond differently to temperature and precipitation changes. In the eastern cordillera, the snowline is presently near the level of the annual 0 ℃ isotherm (4800 m) and at this elevation melting occurs throughout the year making snowlines sensitive to temperature changes but relatively insensitive to accumulation changes. In the western cordillera, the snowline rises 1000 m higher due to increasing aridity. At this higher elevation snowline exhibits a much stronger sensitivity to accumulation changes.

The consistent 1200 m snowline depression along the eastern cordilleras of southern Peru and Bolivia can be modeled by a 5-9 °C cooling which is inconsistent with the rather small (<2 °C) cooling in tropical sea surface temperatures suggested by CLIMAP (1976; 1981). Poor constraints on the timing of the late Pleistocene glacial advance in the central Andes, however, do not allow snowline depressions to unequivocally refute CLIMAP sea surface temperature estimates. The 800-1000 m snowline depression observed in the western cordillera cannot be accounted for solely by cooling, but also requires a precipitation increase.

Introduction

Changes in alpine glaciation are commonly used in paleoclimatic reconstructions to study temperature and precipitation changes as well as to infer moisture sources, wind directions, and effective continentality during the past (e.g. Leonard, 1989; Locke, 1990; Porter, 1977; Porter *et al.*, 1983; Zielinski and McCoy, 1987). Changes in snowline elevation depend on a complex interaction of climatic processes associated with mass accumulation and ablation on a glacier surface. Paleoclimatic interpretations based on snowlines are constrained by difficulties in separating the effects of temperature, precipitation, and net radiation changes. Despite these difficulties, snowline elevation changes between the late Pleistocene and present are one of few methods available for investigating the Last Local Glacial Maximum (LLGM) climate of the central Andes.

Regional snowline studies in the tropical Andes include the work of Nogami (1976) and Satoh (1979). From these studies, which were based on limited data, it is clear that the magnitude of snowline depression ranged from several hundred meters to over 1000 m compared to modern snowline elevations. It is also evident that modern and LLGM snowline elevations show clear regional trends in that both rise from north to south and also rise from the eastern ranges (eastern cordillera) to the western ranges (western cordillera). Recent work (e.g. Fox and Bloom, 1994; Rodbell, 1992; Seltzer, 1987) supported these preliminary findings. However, large discrepancies between various snowline estimates as well as conflicting views about the paleoclimatic interpretation of the snowline changes remain unresolved.

Snowline changes can be used to address two major paleoclimatic questions of interest in the Peruvian-Bolivian Andes. The first concerns the magnitude of tropical temperature depressions during the Last Global Glacial Maximum (LGM). The second, more regional in nature, is whether wetter or drier conditions than present prevailed in Peru and Bolivia during the time of the local glacial advance.

Of particular concern in the tropical Andes, and the tropics in general, is the magnitude of low-latitude temperature depressions during the Last Global Glacial Maximum (LGM). CLIMAP (1976; 1981) reconstructions of sea-surface temperatures (SSTs) in the tropics for the LGM show a deviation from modern temperatures of $< 2°$ C. As pointed out by Rind and Peteet (1985), it is difficult to reconcile this small SST change with the observed snowline depressions and pollen derived temperature estimates in the tropics. In the past several years, a growing body of literature suggests the CLIMAP sea surface temperatures are incorrect. Both oxygen isotope and Sr/Ca thermometry on well dated submerged coral reefs off Barbados Guilderson *et al.* (1994) show that tropical SSTs 18,000-19,000 years ago were 5° C cooler than present. Stute *et al.* (1995) demonstrate a mean annual ground cooling of 5° C at some time prior to 10,000 ^{14}C yr BP by using the concentrations of noble gases dissolved in Brazilian groundwater. At higher elevations, treeline lowering is also considered to be too large to be reconciled with the CLIMAP observations (Hooghiemstra *et al.*, 1992; van der Hammen, 1988). Oxygen and hydrogen isotope changes in ice cores from Nevado Huscarán, Peru, imply a large cooling, possibly as much as 8°- 12° C at 6000 m (Thompson *et al.*, 1995).

Although snowline changes were one of the original lines of evidence used to contradict the original CLIMAP tropical SSTs, they presently do not provide unequivocal evidence to support or refute the CLIMAP SSTs. The major uncertainties associated with snowlines include: (1) age determination of glacial deposits used to infer past snowline position, and (2) uncertainties in the climatic forcings responsible for the observed snowline change. Such uncertainties are clear in the large range of temperature changes in the Peruvian-Bolivian Andes derived from late Pleistocene snowline depression data. These temperature-depression interpretations range from 3.5±1.6° C for the Cordillera Real, Bolivia (Seltzer, 1992) to 10°± 1.9° C in Peru (Fox and Bloom, 1994), as well as numerous intermediate estimates.

Of paleoclimatic interest on a more regional scale is the question of whether the Peruvian-Bolivian Andes experienced wetter or drier conditions during the time of maximum glacial expansion. Snowlines have been used to argue both for wetter conditions (Hastenrath, 1971; Wright, 1983) and drier conditions (Fox and Bloom, 1994; Satoh, 1979). Dramatic changes in the hydrologic balance of the region during the late Pleistocene are also evidenced by significant increases in the volume of lakes on the Altiplano, a large internally drained basin between 14° and 21° S. The largest of these lake level changes could not have been caused by glacial meltwater (Blodgett *et al.*, in press) and have been attributed to increases in precipitation of 50-75% (Hastenrath and Kutzbach, 1985; Kessler, 1984). Blodgett *et al.* (in press) examined the temperature, precipitation, and cloudiness changes that could produce the high lake stands and found that either a 50% precipitation increase, a 10° C cooling, or a doubling of cloudcover could account for the observed lake level change.

The extent to which glacier expansion can add to the debate on tropical SSTs and precipitation on the Altiplano during the Pleistocene depends upon (1) the accurate determination of the magnitude of snowline depression, (2) the relative timing of the maximum glaciation compared to the CLIMAP SST reconstructions, and (3) the climatic interpretation of snowline change. This paper integrates a variety of previous studies with additional mapping of

LLGM features to create a comprehensive map of both modern and LLGM snowline elevations from which to estimate snowline depressions. The growing literature on the behavior of tropical glaciers both from the theoretical standpoint (Kaser, 1995; Kuhn, 1989; Seltzer, 1993) and from field observations (Francou *et al.*, 1995; Hastenrath and Ames, 1995b) provides a physical basis for quantification of the climatic changes reconstructed from observed snowline depressions. The dating of glacial landforms remains poorly constrained and continues to make the comparison of various climatic reconstructions from different proxies difficult.

Study Area

This study encompasses the entire glaciated portion of the Peruvian-Bolivian Andes (**Figure 2.1**). Perennial modern snowcover is found at latitudes south of 8° S, while LLGM glaciation extended northward to 5° S. At these latitudes, the Andes are divided into the eastern and western cordilleras. The cordilleras in southern Peru and Bolivia are separated by a large intermontane internally drained basin of high elevation but low relief known as the Altiplano, which is second only to the Tibetan Plateau in height and aerial extent. The present day topography of the tropical Andes (**Figure 2.1)** is a primary tectonic signal of late Cenozoic mountain building processes in this non-collisional mountain system (Isacks, 1988). Tectonic deformation accompanying plateau formation in the region is thought to have occurred in

Figure 2.1: Topography of the central Andes shown in grayscale (Defense Mapping Agency, 1989; Isacks, 1988). Locations of sites for which mimimum limiting dates for the Last Glacial Maximum have been aquired are also shown: (A) Laguan Junín, (B) Cordillera Vilcanota, (C) Laguna Kollpa Kkota.

two major phases. The first major phase occurred in the late Miocene and resulted in the compressional deformation of the Altiplano Plateau. During the second major phase, which occurred after 10 Ma, deformation migrated eastward to the Subandean fold and thrust belt (Gubbels *et al.*, 1993). This deformation allowed mountains within the region to achieve an elevation sufficient to be glaciated by the late Pliocene, at least in Bolivia, as evidenced by the Calvario Drift, which has an age of 2.2 myr (Thouveny and Servant, 1989).

Denudation rates have been determined for only one site within the region, the Zongo Pluton, Cordillera Real, Bolivia. Masek *et al.* (1994) determined that denudation rates over the last 20 myr range between 0.2 and 0.9 mm a^{-1} based on fission track studies of Benjamin *et al.* (1987). Even if the topography is rising at these rates, topographic changes due to tectonic processes over the last glacial cycle (~100,000 years) would amount to less than 100 meters. Thus, it is unlikely that tectonic processes have significantly altered regional snowline patterns.

Glacial Chronology

Almost all of the dates for the late-Pleistocene glacial chronology in Peru and Bolivia are minimum-limiting ages for deglaciation (Seltzer, 1990). Lakes and peatlands have been cored, and basal organic-rich material has been dated by conventional radiocarbon methods. In general, most lakes and peatlands formed after the last deglaciation, which began more than 13,000 ^{14}C yr BP.

At several sites in the tropical Andes older dates have been obtained from lakes and peatlands, and these dates imply that the LLGM may predate 20,000 ^{14}C yr BP. The stratigraphy from a 30 m core from Laguna Junín indicates that glaciers were in the vicinity of the lake from 24,000 to 12,000 ^{14}C yr BP (Hansen *et al.*, 1984). However, radiocarbon dates on the Junin Plain indicate that the LLGM may predate 42,000 ^{14}C yr BP (Wright, 1983). At the Cordillera Vilcanota, a peat was that was overridden by a glacier about 14,000 ^{14}C yr BP, but another peatland near the glacier has numerous basal dates that imply the glacial maximum in the area occurred prior to 28,000 ^{14}C yr BP (Mercer, 1984). At Laguna Kollpa Kkota in the Eastern cordillera, Bolivia (Seltzer, 1994), and at Laguna Ciega, Sierra Nevada del Cocuy, Colombia (van der Hammen *et al.*, 1981), basal radiocarbon dates imply that the LLGM occurred more than 20,000 ^{14}C yr BP. On the Ruíz-Tolima massif in the Cordillera Central, Columbia, glaciers achieved their greatest extent when the climate was cold and moist, corresponding to oxygen-isotope stage three. Glaciers were still expanding in the period 28,000 to 21,000 yr BP, but shrank considerably after this time (Thouret *et al.*, 1996).

There remains considerable uncertainty about the timing of the LLGM in the tropical Andes because of the lack of maximum limiting ages for the glacial advances. This is not an unusual circumstance for alpine regions because of the relative lack of datable material in association with moraines (cf. Gillespie and Molnar, 1995). Thus, it remains problematic whether or not the local glacial maximum in the central Andes was contemporaneous with the timing of LLGM which, based on upon the maximum depletion of $\delta^{18}O$ in marine carbonates for the last glacial cycle (~100,000 yrs), occurred from 23,000 to 14,000 yr BP (CLIMAP, 1981). Although it appears that the maximum glacial extent in the central Andes preceded the last global glacial maximum, the question remains unresolved.

Regional Snowline Mapping

Introduction

Presented here is an effort to map perennial snow and glaciers and the maximum limit of glacier extent reached during the LLGM at a regional scale. This work represents an integration of previous regional scale mapping (Fox, 1993; Fox and Bloom, 1994; Seltzer, 1987) with considerable original mapping. The problem is approached in traditional fashion by determining the elevation of both modern and LLGM snowlines and calculate the snowline depression as the difference between the two elevations. The primary difference between this study and previous work is the use of remote sensing and geographic information system technologies to produce a comprehensive regional overview, rather than a detailed picture at single sites.

In this study, the term *snowline* is adopted as an approximation of the glacier equilibrium line altitude (ELA) both at present and during the past. For the modern case, the lower limit of perennial snowcover or glaciers is used to define the modern ELA. Although this definition varies from the classic definition of orographic snowline as the lower limit of perennial snow patches (Andrews, 1975), it is necessary because it is difficult to distinguish snow patches from small glaciers on topographic maps and satellite imagery in the central Andes. While this snowline lies below the glacial ELA, altitude trends in the two do parallel each other (Andrews, 1975).

Over large portions of the region, no glacier inventory exists. This precludes using either glacial ELAs or the median glacier elevation to define the snowline. While the snowline used in this study is not valid for modern glacial studies or at small spatial scales, it is reasonable for a regional scale paleoclimatic study. The errors introduced by using this approximation as opposed to other glaciological measures such as estimates of glacial ELAs are discussed below.

To create regional scale maps of LLGM snowline elevation two simplifying assumptions are necessary. First, we assume the geomorphic features used to define the state of glaciation formed contemporaneously during what we refer to as the LLGM. While this assumption may not be true, it is reasonable given the paucity of dates constraining the LLGM (discussed above). Second, two methodologies (cirque floors in Peru, and mapped moraines and a toe-to-headwall altitude ratio south of 15° S) were used to estimate the ELAs of the late Pleistocene glaciers. This necessitates the assumption that all methods yield equivalent snowline elevations. The validity of these assumptions is discussed below.

Construction of regional snowline maps requires determination of a representative snowline elevation for an area based on the individual measurements of modern and LLGM snowline elevations. Modern and LLGM snowline estimates were calculated for a 15´ x 15´ (approximately 28 km x 28 km) series of cells covering the study area. This cell size is small enough to capture some relatively small scale features, especially obvious for the LLGM case, but large enough to provide a useful regional view.

Modern Snowline

To produce a comprehensive spatial map of modern snowline altitude in the central Andes it was necessary to integrate snowline elevations derived by two methodologies. In Peru, modern snowline elevations from the study by Fox and Bloom (1994) were used. In that study, the modern snowline was taken as the lower limit of snowcover recorded on the Peruvian 1:100,000 topographic maps published by the IGM in Lima, Peru. Snowline elevations were recorded every 1-2 km along the lower limit of snowcover to capture the local variability in snowline altitudes (Fox and Bloom, 1994).

In southern Peru, Bolivia, and northern Chile, Landsat Thematic Mapper (TM) imagery was employed to map the lower limit of snowcover. Most of the images employed for this study were acquired during the dry seasons (June-August) of 1984-1987. Daily temperatures above freezing, even at high elevations, quickly melt snow falling on soil or bare rock (Ribstein *et al.*, 1995). Dry conditions combined with rapid melting of freshly fallen snow make the snowcover captured on the satellite imagery close to its minimum extent. Nevertheless some interannual variability in snowcover can be seen in areas in where multitemporal TM coverage is available. This variability adds uncertainty to the estimates of modern snowline elevation. In cases where snowcover was clearly non-permanent (e.g. confined to narrow stream valleys), the elevations were discarded. However, the modern snowlines elevations are estimates of the lowest elevation that modern snowline reaches and are the greatest source of error in this analysis.

An algorithm developed by Dozier (1989) was used to determine which image pixels were snow covered and proved highly accurate in the arid tropical Andes. The only natural surface misclassfied as snow were the edges of the *Salars* (salt pans) of Copasia and Uyuni, which were covered with a thin layer of water. However, these areas are easily removed from further analysis.

The images were preprocessed to allow comparisons of TM images acquired at different times and varying atmospheric conditions. The original Digital Number (DN) values of the TM image were converted to at-satellite reflectance (Markum and Barker, 1986). This procedure corrects for time-dependent changes in the satellite instrumentation and for seasonal changes in illumination. A simple atmospheric correction was applied using the modified blackbody method of Chavez (1988). This preprocessing allowed for a uniform binary classification criteria to be applied to determine if a pixel was snow covered. The classification criteria are as follows: reflectance in TM band 1 (0.45-0.52 μm) greater than 15%; reflectance in TM band 5 (1.55-1.75 μm) less than 15%; and the normalized difference ratio between TM band 2 (0.52-0.60 μm) and TM band 5, [(TM2-TM5)/(TM2+TM5)], greater than 0.50. An example of this classification is shown for Nevado Sajama (**Figure 2.2**).

Once the snowcover classification was completed, the snowcover maps were transformed into the UTM coordinate system using a first or second order polynomial ground-control point warp. For the tropical Andes, the root mean square (RMS) errors in the horizontal for the control points were better than 2-2.5 pixels (60-75 m). Given the paucity of cultural features to provide good ground control, these RMS errors were considered acceptable. These geocoded snowcover images were then integrated with topographic maps or digital elevation models (DEMs) derived from these topographic maps produced the Instituto Geografia Militar in Peru or Bolivia at scales ranging from 1:50,000 to 1:250,000. Snowline elevation was recorded in a manner analogous to that applied in Peru.

For the modern case, the estimate of regional snowline elevation for each cell was taken to be the mean value of all individual snowline estimates falling within each 15´ x 15´ cell. The mean elevation, rather than the lowest

Figure 2.2: Example of computer snowcover classification and Last Glacial Maximum snowcover mapping for Nevado Sajama, Bolivia (S 18 º07´, W 68 º 52´). (A) Landsat TM (Band 4) image of the region, (B) Dozier (1989) snowcover classification for the image, and (C) Last Glacial Maximum glacial extent for the region mapped from Landsat TM imagery.

elevation, was selected for several reasons. The major problem with the estimate of modern snowline is that it may not represent the minimum snow extent. By using a mean rather than a minimum snowline elevation for each cell, the problem of excessively low snowline elevations is minimized. In many cases only a small number of modern snowcover elevations may be present within a cell and

may represent snowcover at more than one time. By using a mean value, the temporal variability inherent in modern snowcover is muted.

It is possible to examine the differences in snowline elevations determined from topographic maps and from Landsat TM for an area in southern Peru where both methods were used (**Table 2.1**). The comparisons were accomplished by calculating the regional snowline for each cell, as described above, and comparing the differences between methods. The differences between snowline elevations using these techniques is small, with a mean difference of only 90 m. More importantly, it is possible to quantify the effect of using the lower limit of perennial snowcover or glaciers rather than a median glacier elevation (or transient snowline) to estimate the ELA. This is possible by comparing snowlines elevations derived from Landsat TM with median glacier elevations and transient snowline estimates from the glacier inventory compiled by Jordan (1991) for a portion in the eastern cordilleras of Bolivia where both measurements where made. The mean difference between these snowline estimates and the glaciological estimates of the ELA are small, on a regional scale only 146 m (**Table 2.1**). This

Table 2.1: Modern snowline comparisons in Peru and Bolivia. Klein *et al.* (This study) used Landsat TM imagery to determine modern snowlines, Jordan (1991) used aerial photographs, and Fox and Bloom (1994) used the Peruvian 1:100,000 topographic maps.

	Mean (m)	Standard Deviation (m)	Minimum (m)	Maximum (m)	Number
Klein *et al*. minus Jordan	-146	60	-243	-70	13
Klein *et al*. minus Fox and Bloom	90	110	-179	330	38

difference is approximately what the expected elevational difference between the lower limit of the glaciers (considered as the lower limit of snowcover in the TM analysis) and the median elevation of small Bolivian glaciers would be.

Last Local Glacial Maximum (LLGM) Snowline

As with modern snowline, two methods were used to determine the regional LLGM snowline surface in the tropical Andes. In Peru where comprehensive TM coverage was not available to us, cirque floor altitudes were used as an approximation of the regional snowline. The regional snowline altitudes here are based on the work of Fox and Bloom (1994) which provided more complete coverage of Peru than did the earlier study by Seltzer (1987). Both studies were based on careful analysis of identical topographic maps and elevational differences in the estimated LGM snowline can be directly attributed to bias in cirque selection. The two studies were compared to assess the effect of observational biases on the estimated LGM snowline and the results are discussed below.

In southern Peru, Bolivia, and northern Chile where TM coverage was used, the former extent of each glacier was mapped from TM images by identifying and mapping the position of terminal moraines. An example of the mapping is shown in **Figure 2.2**. In a few cases abrupt changes in valley form or the presence of a glacial cirque were used where moraines forms were not clearly visible. The use of digital TM images allowed mapping at scales of 1:50,000 or better. In addition to the TM, both SPOT images and aerial photographs were used to check and improve mapping in the Bolivian Cordillera Real. The use of these higher spatial resolution data, however, did not significantly alter the original TM glacial mapping and served as a validation of the TM mapping.

Once the former extent of each glacier was mapped, the snowline altitude of the glacier was calculated. In this study the LLGM snowline altitude was approximated using a "toe-to-headwall altitude ratio" (THAR) method (Meierding, 1982). The toe (minimum) and headwall (maximum) elevations for each former glacier were determined and the ELA altitude calculated as:

$$\text{ELA} = \text{toe} + (\text{THAR})(\text{headwall-toe}) \tag{1}$$

Several factors were considered in selecting a THAR value. First, THAR and AAR (Accumulation Area Ratio) values for modern glaciers in the region were evaluated (T. Blodgett, unpublished data) based on the Bolivian Glacial Inventory by Jordan (1991). For 78 modern glaciers, the average THAR was 0.35 and the average AAR was 0.67 (**Figure 2.3**). However, it is not clear whether the modern AAR and THAR values are appropriate for these late Pleistocene glaciers. One conspicuous feature of these former glaciers is their long narrow tongues which are not general feature of the region's modern glaciers. If the hypsometries (area-altitude relationship) of modern and late Pleistocene glaciers do differ, then application of a modern THAR value to

Figure 2.3: Histograms of **(A)** toe-to headwall altitude ratios (THARs) and **(B)** accumulation area ratios (AARs) calculated for seventy-eight modern Bolivian glaciers based on the glacier mapping by Jordan (1991).

the paleo-glaciers will cause the estimated AARs of paleo-glaciers to deviate considerably from observed values. To assure that AAR ratios of the paleo-glaciers are close to the modern average, it is necessary to adjust the THAR value.

To accomplish this, the relationship between modern THARs and AARs was investigated for a subset of sixty-nine late Pleistocene glaciers on three Bolivian massifs: Cerro Tunupa, Cerro Cosuña, and Kari Kari. The surfaces of these glaciers were reconstructed by Fox (1993)following the methods developed by Porter (1975). For each of the reconstructed glaciers, a hypsometric curve (**Figure 2.4A**) was calculated and the relationship between AAR and THAR determined (**Figure 2.4B**). For the sixty-nine reconstructed LGM glaciers an average THAR between 0.40-0.45 provided the best AAR match with present AAR. Therefore, a THAR of 0.45 was used to calculate the LLGM snowline for each mapped glacier. Like their modern counterparts, most of the paleo-glaciers were quite small (median area = 0.87 km^2) and had a limited elevational range (median elevation range = 400 m). Thus the estimated paleo-ELA is fairly insensitive to the exact choice of a THAR. For the reconstructed glaciers, the mean elevational difference in the estimated ELA for a THAR change 0.05 is only 38 meters. The minimum (toe) and maximum (headwall) elevation necessary for the calculation of a THAR each of the paleo-glaciers were determined by overlaying the glacial extents on 1:50,000 and 1:250,000 topographic maps (produced by the

Figure 2.4: **(A)** A hypsometric curve of the reconstructed surface of a late Pleistocene glacier on Kari Kari. **(B)** the relationship between the calculated toe-to-headwall altitude ratio (THAR) and accumulation area ratio (AAR) calculated from the hypsometric curve in **(A).**

Instituto Geografia Militar, La Paz, Bolivia) and recording the minimum and maximum elevations.

The selected THAR value agrees well with published values. Meierding (1982) found a THAR of 0.35-0.40 to give the best results for former glaciers in the Rockies of the western United States. In the Pacific Northwest a THAR of 0.50 has been used by several workers (Porter, 1964; Scott, 1977) and for modern glaciers in Alaska, Péwé and Reger (1972) found that 0.66 was the most appropriate THAR

The LLGM snowline estimates calculated for each paleoglacier, were used to create a regional LLGM snowline. The minimum ELA estimate within each 15´x 15´ cell was selected. Unlike the modern case, each cell may contain hundreds of paleo-snowline elevations. Some of these individual elevations definitely do not represent the minimum elevation to which glaciers descended during LLGM and use of a mean value was considered inappropriate. It could be argued that an average of several of the lowest snowline elevations should be used instead of the single lowest value, but upon examination, averaging of snowline values had little effect on the calculated regional snowline estimates.

Two methodologies were employed in construction of the LLGM snowline map: cirque floor altitudes in Peru and glacial reconstructions in southern Peru, Bolivia, and Northern Chile. It is quite possible that the ELA estimates may differ as a function of the methodology. In southern Peru where both methodologies were employed, a comparison of methods is possible. However, because of limitations in the spatial accuracy of one of the comparison datasets (Seltzer, 1987) a 30´ x 30´ rather than 15´ x 15´ averaging cell was used.

While comparisons between snowline elevation estimates derived from various techniques have been undertaken previously (e.g. Meierding, 1982), few studies have examined differences in snowline altitudes between studies that employed the same technique in the same area. Such a comparison is possible in Peru where two independent studies, Fox and Bloom (1994) and Seltzer (1987), mapped cirque floor altitudes using identical techniques and maps. Differences in snowline elevation between the two studies can be attributable primarily to varying cirque selection between the studies. For the areas of overlap, the mean difference in regional snowline estimates on 30´ x 30´ cells was nearly zero. However, a surprising number of localities exist where the differences were quite high ($\geq$ 450 m). Differences in snowline altitude of $\geq$ 100 meters occurred in 45% of the cells, while differences of $\geq$ 200 occurred at 18% of the cells. These results highlight the uncertainty in regional estimates of LLGM snowline, for cirque floors at least, due solely to observational bias. The differences between LLGM snowline mappings are shown in **Table 2.2** and **Figure 2.5.**

Differences in cirque floor elevations and snowlines derived from reconstructed glaciers, were also compared. The mean difference for the entire area of overlap is 170 m. Two factors can be invoked to explain the

Table 2.2: LLGM snowline difference estimates for Peru: a comparison of methods. Fox and Bloom (1994) and Seltzer (1987) used cirque floor altitudes and Klein *et al.* (This study) used the THAR methodology with LANDSAT TM mapping of glacial moraines to determine the LGM snowlines.

	Mean (m)	Standard Deviation (m)	Minimum (m)	Maximum (m)	Number
Fox and Bloom minus Seltzer	-6	134	-300	450	21
Fox and Bloom minus Klein *et al.*	173	243	-160	760	20
Seltzer minus Klein *et al.*	80	254	-370	760	15

Figure 2.5: Differences in LLGM snowline elevation estimates in Peru for three studies. Top three panels show elevational differences in snowline estimates based on cirque floors (Fox and Bloom, 1994; Seltzer, 1987) and glacier reconstructions (Klein et al. this study). Each cell in represents a 0.5 º x 0.5 ºarea. The lower three panels are histograms of the differences shown above. (A) Fox and Bloom minus Seltzer, (B) Fox and Bloom minus Klein et al., (C) Seltzer minus Klein et al.

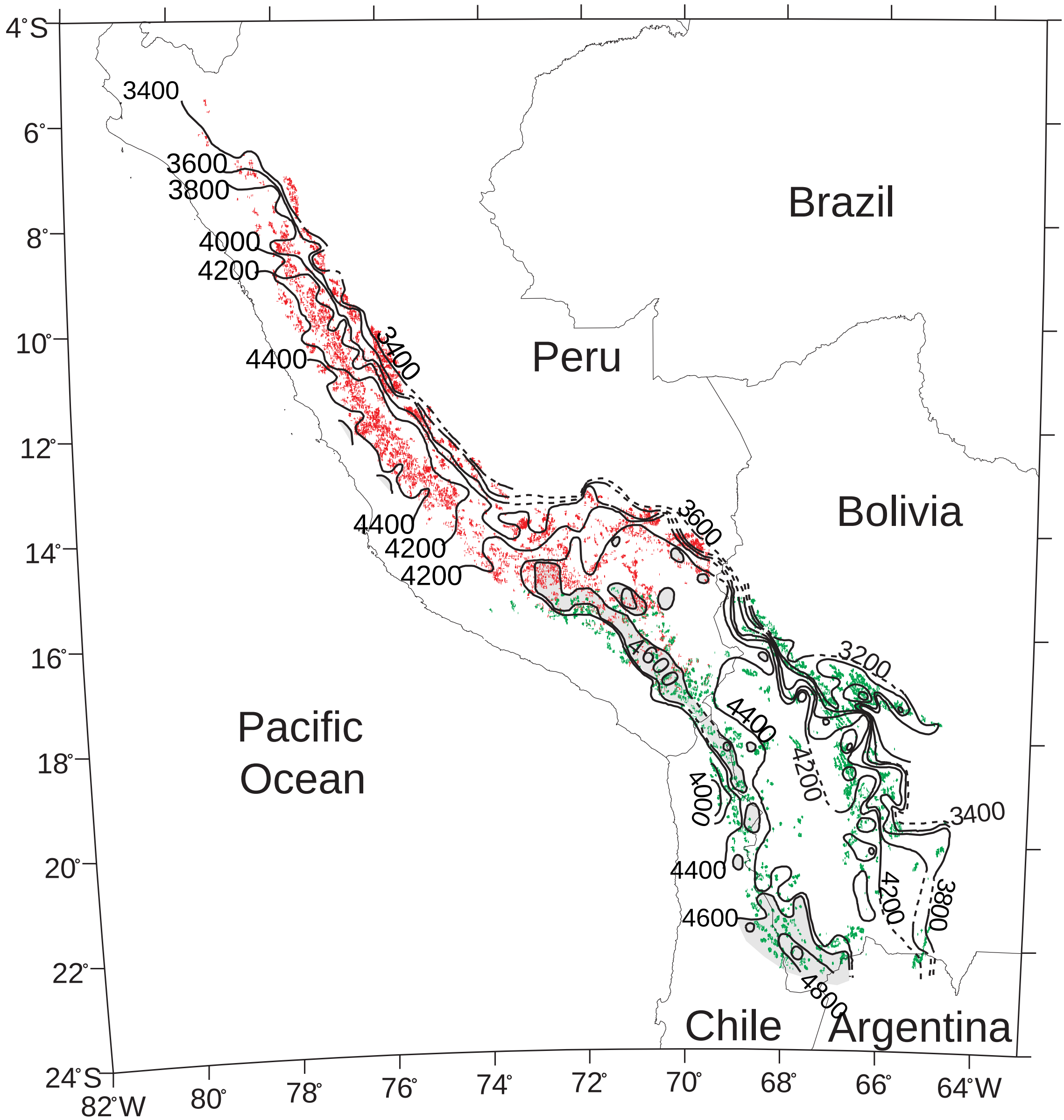
difference. First, for each individual paleo-glacier the two methods may produce different snowline estimates. Secondly, in the Peruvian Andes the two methodologies spatially sample different portions of the landscape within each cell. Some glaciated areas are favorable for moraine preservations and some areas for cirque development. These areas may or may not be coincident. This is especially apparent in southwestern Peru where cirques are seldom found on the western side of the western cordilleras. Here it is possible to observe moraines in the TM imagery where no cirques are present. In these areas the snowline estimates from the TM mapping are lower than from the cirque estimates and is the major cause for the differences in snowline estimates between techniques. It also is responsible for the large (700 m) elevation difference observed in one cell. As with the modern case, differences in the LLGM snowline elevation at a regional scale do result from employing different methods, but do not preclude integrating the various methods to produce regional snowline maps.

Last Local Glacial Maximum Snowline depression

The contour maps for the modern and LLGM cases (**Figures 2.6** and **2.7**, respectively), were created from the resulting regional values by hand contouring. From these maps a triangular irregular network (TIN) algorithm (Peucker *et al.*, 1978) was used to create continuous modern and LLGM snowline surfaces of the entire region. Snowline depression (**Figure 2.8**) for

Figure 2.6: Modern snowline elevations for the central Andes. Contour interval is meters and contours are dashed where approximated. Each point represents a snowline elevation observation

Figure 2.7: Last Glacial Maximum snowline elevations for the central Andes. Contour interval is 200 meters and contours are dashed where approximated. Each point represents a snowline elevation observation



4°S
6°
8°
10°
12°
14°
16°
18°
20°
22°
24°S
82°W
80°
78°
76°
74°
72°
70°
68°
66°
64°W
Brazil
Peru
Bolivia
Pacific
Ocean
Chile
Argentina
3400
3600
3800
4000
4200
4400
4600
4800
3200

Figure 2.8: Last Local Glacial Maximum snowline depression. The depression is calculated by subtracting the LLGM snowline surface (**Figure 2.6**) from the modern snowline surface (**Figure 2.7**). Contour interval is 100 m.

the region was calculated by subtracting the LLGM surface from the modern surface and the resulting values were then contoured.

The overall error in estimation of snowline depression includes elevation errors in both the modern and LLGM surfaces. The various techniques used to estimate both the modern and LLGM snowline produce snowline estimates that are on average within 200 meters of each other. The LLGM surface contains additional uncertainties due to subjective selection of cirques, and quite possibly moraines. In the contouring process, much of the high frequency variability both within and between the input sources for both surfaces is smoothed out. Given all these uncertainties, the accuracy of snowline depression at any one point is probably on the order of ± 100 - 200 m. However, the consistent regional trends of most interest here are undoubtedly real.

Snowline Trends

The major feature of the modern snowline in the tropical Andes is its general northeast-southwest rise across the region (**Figure 2.6**). The lowest measured snowline elevations (4400 m) are found in the eastern cordillera of Bolivia. Low modern snowlines, at or below 4700 m, continue northward throughout the length of the Peruvian Andes. In central Peru, the snowline rises westward to 5100 m. In Bolivia, the snowline rises quickly from 4400 meters to 5000 along narrow ranges comprising the eastern cordillera. The snowline rises gradually southwest across the Altiplano to extremely high elevations, over 5800 meters, in the western cordillera along the Bolivia-Chile border.

The general configuration of the snowline during LLGM (**Figure 2.7**) was similar to present. The LLGM snowline shows both an east-west and north-south rise across the tropical Andes similar to the modern case (**Figure 2.6**). Along the eastern cordillera, the LLGM snowline descended to elevations as low as 3200 m and consistently was at or below 3400 along nearly the entire length of the eastern cordillera of Peru and Bolivia. In central Peru, the LLGM snowline rose westward to elevations >4400 m. In Bolivia, the LLGM snowline rose steeply across the eastern cordillera then gently rose southwest across the plateau to >4800 in the western cordillera. In southwestern Peru, and northern Chile, the LLGM snowline shows a slight descent to lower elevations along the western slopes of the western cordillera (**Figure 2.9 transect C**).

These results generally support the earlier interpretations of Satoh (1979) but with much greater detail. One important detail is that both the modern and LLGM snowline rise steeply across the eastern cordillera. These cordillera form an effective moisture barrier and the steep rise in snowline is in response to a strong gradient in precipitation. In areas where the cordillera are higher and hence form a more effective orographic barrier the modern and LLGM east-west snowline gradients are steeper.

Several authors (Fox and Bloom, 1994; Satoh, 1979; Seltzer, 1987) have noted an increased east-west snowline gradient for the LLGM. This has

Figure 2.9: Snowline transects across the central Andes. Each transect shows the average elevation of the modern and LLGM snowlines within a 50 km wide x 10 km long swath. Average snowline depression and precipitation (in m) for each swath are also shown. The location of the six transects are illustrated on the inset map. The gray background represents the range of topographic elevations found within the 50 km transect. East-west transects (**A-D**) and north-south transects (**E & F**) are shown.

been considered indicative of drier conditions. However, modern and LLGM snowline gradients are not uniform over the region, nor is there a uniform gradient increase. The sparse spatial sampling of the modern snowline in many areas makes the use of gradient changes as a paleoclimatic indicator uncertain (e.g. **Figure 2.9 transect C**).

Snowline during the LLGM was depressed 500-1200 m compared to present (**Figure 2.8**) and was greatest along the eastern cordillera of Peru and Bolivia. A consistent 1200 m depression in snowline existed along the length of the Peruvian Andes and southward to ~17 ºS in Bolivia. Farther south along the eastern Andean margin, snowline depression equaled or exceeded 1000 m. A small area of snowline depression in excess of 1000 m is found in northern Chile. However, uncertainties in the modern snowline elevation limit the certainty of the exact magnitude.

The snowline maps presented here is consistent with those of previous studies (Fox, 1993; Fox and Bloom, 1994). The notable exception is that the snowline depression along the eastern cordillera is 100 to 200 meters less in this study than in the previous studies (1200 versus 1400 m). This difference appears primarily in the snowline depression map and is not as obvious in either the modern or LGM snowline. It is possibly explained by differences arising in creating the snowline surfaces from the contours. It should be noted that these differences lie at the edges of the grids, so care should be taken in their interpretation. Given uncertainties in any surface creation algorithm, the more conservative (1200 m) depression seems warranted.

The magnitude of snowline depression over much of the tropical Andes was <1000 m, yet large areas of southeast Peru and Bolivia experienced only a 500 to 800 m snowline shift. These areas of diminished snowline depression correspond to areas of the Altiplano characterized by high elevation but low relief (**Figure 2.1**). This coincidence suggests the presence of a high plateau limits the elevation to which glaciers could descend. The potential thus exists that a non-climatic factor may control the elevational limit of LLGM glacial expansion in portions of the tropical Andes. As glaciers descended from the mountains onto the upper elevations of the Altiplano, they were unable to descend lower without experiencing considerable lateral expansion. The coincidence of multiple phases of glaciation reaching nearly the same elevation along the western slope of the Cordillera Real, Bolivia (Argollo, 1980; Gouze *et al.*, 1986; Seltzer, 1992) implies that either the region experienced very similar climate changes several times in the past or there is an non-climatic threshold in the glacier system.

The 1000 m of snowline depression often assumed for the tropics (Brocker and Denton, 1989) is a dramatic oversimplification of the actual depression observed in the tropical Andes. While certain areas of the tropical Andes, in particular the eastern slopes of the eastern cordillera, certainly did experience a snowline lowering of 1000 m ± 200 m, this value cannot be considered typical of large areas of the Peruvian and Bolivian Andes. As the limit of glacier expansion may be controlled by non-climatic factors, such as the varying hypsometry of glaciers as they expand downvalley, care must be taken when selecting values of snowline depression for paleoclimatic interpretation.

Paleoclimate Analysis

Introduction and Model Description

Because the elevation of the snowline, or more correctly, the equilibrium line altitude (ELA) is a complex function of accumulation and ablation processes, it is impossible to determine uniquely the climatic changes responsible for the observed LLGM snowline depression in the central Andes. Here the problem of climatic interpretation is approached by using a theoretical model of tropical glacier behavior developed by Kaser (1995) to estimate the response of the ELA to climatic perturbations. Concerted efforts, described below, are made to portray the glaciologically important aspects of tropical Andean climate adequately. The effort reveal a plausible set of climatic change scenarios consistent with the observed ELA depression. However, application of glaciological models to paleoclimatic analysis requires caution. A fundamental assumption made here is that climatically important parameters for the period of interest, such as the lapse rate and vertical accumulation gradient, were similar to modern conditions. Air temperature, precipitation, and the duration of melting are assumed to vary linearly as a function of elevation and the ELA response is modeled as linear sum of all climatic perturbations.

The basis of the model is the vertical mass balance profile. At any elevation, the mass balance (b) is the difference between accumulation (c) and ablation (a).

$$b = c - a \tag{2}$$

The model creates a mass balance profile using linear vertical profiles of accumulation and ablation representative of observed conditions as inputs. The response of the ELA to a climatic change is determined by calculating two mass balance profiles, the first under unperturbed conditions and the second with a climatic perturbation (e.g. air temperature increased by 1° C). The response of the ELA to this climatic perturbation is the difference in the elevation of the equilibrium line (b=0) between the two states.

Accumulation is the sum of all processes that add mass to a given point on a glacier such as snowfall, avalanching and wind-drifting. Ablation is calculated as the sum of the energy available for melting from net radiation (Q_r) and sensible heat transfer during the time (τ_m) when the glacier surface is melting. Sensible heat transfer is simply the difference between the air (T_a) and surface (T_s) temperature, which is assumed to be 0° C during melting, multiplied by a bulk transfer coefficient for sensible heat (α). In this simple model, all meltwater is assumed to exit from the glacier and refreezing is neglected. Thus the ablation (a) is given by:

$$a = \frac{t_m}{L_m}\left[Q_r + \alpha(T_a - T_s)\right] \qquad (3)$$

where L_m is the latent heat of fusion (3.34 x 10^5 J kg^{-1}). Latent heat transfer, which in most cases is an order of magnitude less than either net radiation or sensible heat flux (Paterson, 1994) is neglected. Sublimation, thought to be an important ablation process in the arid western Cordilleras of the tropical Andes (Seltzer, 1993) is also neglected. Where air temperatures are above 0° C, the specific balance at any elevation on the glacier can then be calculated as:

$$b = c - \frac{t_m}{L_m}\left[Q_r + \alpha(T_a - T_s)\right] \qquad (4)$$

When air temperatures are below freezing the net balance is simply equal to accumulation (b=c). This model was used in conjunction with vertical profiles of accumulation, air temperature, net radiation, and duration of melt to create synthetic mass balance profiles.

Glacialogically important aspects of central Andean climate

Application of the mass balance model to central Andean conditions is limited by the dearth of climate observations, especially at high elevations, and a lack of understanding of how these tropical glaciers respond to climatic changes. To overcome limitations in ground station climatologies, vertical profiles of air temperature, net radiation, specific humidity and duration of melt were calculated from the Goddard Earth Observing System Office

(GAO) General Circulation Model (GEOS-1) that is available for the years 1986-1990. GEOS-1 is a multi-year global atmospheric data set on a 2.5° latitude x 2.0° longitude grid with 18 vertical levels in the atmosphere. The GEOS-1 model uses a data assimilation scheme to incorporate ground-based and satellite observations in a general circulation model to produce a climatology (Schubert *et al.*, 1993). While the assimilated model output may deviate slightly from observations, the averaged fields employed here should be quite reasonable representations of climatic conditions in the region's free atmosphere.

Although tropical glaciers respond differently to climate changes than their better studied mid-latitude equivalents, our understanding is limited by the small number of field investigations. The only long term glaciological study in the tropics has been done on Lewis Glacier on Mt. Kenya (Hastenrath, 1984; Hastenrath, 1989). In the central Andes only a few glaciological studies have been undertaken. The Quelccaya Ice Cap, Peru, has been the site of both glaciological and ice core research (Hastenrath, 1978; Thompson, 1979; Thompson *et al.*, 1995). Two glaciers in the tropical Andes: Yamamarey Glacier, Peru (Hastenrath and Ames, 1995a; Hastenrath and Ames, 1995b; Kaser *et al.*, 1990), and the Zongo Glacier in Bolivia (Francou *et al.*, 1995; Ribstein *et al.*, 1995), have received recent attention. While many questions remain, a basic understanding of the effect of climate on the glaciers of the tropical Andes is beginning to emerge.

These field studies serve as the basis for a theoretical model of tropical glacier behavior developed by Kaser (1995). His modeling indicates that glacier behavior is strongly influenced by two characteristics of tropical climate: small temperature differences throughout the year and the seasonality of precipitation. In addition, the vertical precipitation distribution in the tropics varies greatly from that in many mid-latitude mountain ranges. These factors combine to make glaciers in the humid tropics more sensitive to temperature changes and relatively insensitive to radiation and precipitation changes. In the dry tropics the sensitivity of glaciers to precipitation changes increases.

The temperature regime of the central Andes is characterized by small thermal seasonality and a predominance of the diurnal cycle as is illustrated by differences between average monthly minimum and maximum temperatures exceeding the range of mean monthly temperatures (**Figure 2.10**). This thermal regime has two important implications for glacier behavior. First, the lack of appreciable thermal seasonality means ablation may not be restricted to a single season, but can occur throughout the year, while the dominance of the diurnal cycle limits ablation to a portion of the day. Second, the lack of thermal seasonality leads to a relatively constant elevation of the 0° C isotherm throughout the year. This makes the annual position of the freezing level a fundamental control on ablation. Below the elevation of the annual 0° C level ablation will be quite intense, while slightly above it ablation will be much reduced as air temperatures seldom

Figure 2.10: Monthly temperatures and precipitation for high elevation Andean stations (1977-1991). For each station, the mean monthly temperature (°C) is represented by the solid square. Averages minimum and maximum temperatures for each month surround the mean. Mean monthly precipitation (mm) is shown in bar graph form. (data is taken from Global Daily Summaries, National Climatic Data Center, Asheville, NC)

will exceed freezing. Previous research has demonstrated that in the tropics, ELA sensitivity is strongly dependent on the position of the ELA relative to the position of the 0° C isotherm. An ELA lying at or below the level of the 0 º C isotherm is primarily sensitive to temperature changes whereas an ELA lying above the level of the 0 ℃ isotherm is much more sensitive to accumulation changes (Kaser, 1995; Kuhn, 1989). The elevational position of the ELA relative to the 0° C isotherm is dictated by amount of accumulation. As accumulation decreases, modern snowlines are observed to rise well above the annual freezing level. In this model the strong dependence of ablation on the freezing level is captured by profiles of air temperature (T_a) profiles and duration of melt (τ_m).

A time series of the elevation of the 0° C isotherm was developed from the GEOS-1 data for 1989. The four hour timestep of the GEOS-1 model captures the strong diurnal temperature cycle of the tropics. At each timestep the elevation of the 0° C isotherm is determined and the cumulative time each elevation spends above freezing is calculated. As is seen in **Figure 2.11A**, the 0° C isotherm is higher and the vertical gradient of the duration of melt (t_m) is steeper in the tropics than in the mid-latitudes. In this model, melt duration is schematically represented by assuming it varies linearly with elevation from 0 to 360 days. Melting is assumed to occur one-half the time at the elevation of the annual 0° C where τ_m is fixed at 180 days.

Figure 2.11: (A) Isotherm elevations for the central Andes (solid) and Austrian Alps (dashed) calculated from the GEOS-1 Model. The light lines represent the calculated elevation of the 0 °C isotherm at 3 hour intervals for the year 1989. The two heavy lines represent the amount of time (in days) air temperature is above freezing at a given elevation for the same period. The dashed line is the elevation of the annual average 0° C isotherm. **(B)** Compilation of 0 °C isotherm estimates. Shown as solid (January) and dashed (July) lines are meridional transects at 70 °longitude from Crutcher *et al.* (1971). Shown as short lines are summer (solid) and winter (dashed) radiosonde estimates for Lima and Antofagasta (Schwerdtfeger, 1976) as well as winter estimates for La Paz (unpublished data). Estimates of the annual 0° C isotherm elevation from ground station climatologies (see text for description) are shown as solid boxes (annual elevation). The topographic envelope of the central Andes is shown in gray.

The elevation of the 0° C isotherm calculated from the GEOS-1 data is consistent with that from radiosonde and ground stations (**Figure 2.11B**). In Peru, the annual mean elevation of the 0° isotherm in the free atmosphere from radiosonde data is estimated to be between 4800 and 5000 meters (Hastenrath, 1967; Satoh, 1979) which compares well to estimates of 4900 to 5000 meters based on ground station climatologies (Fox and Bloom, 1994; Johnson, 1976; Seltzer, 1987). Farther south in Bolivia the mean annual elevation is slightly lower at 4600 to 4800 meters (Fox, 1993; Graf, 1981; Hastenrath and Kutzbach, 1985; Satoh, 1979). Seasonal variation in the elevation of the 0° C isotherm does exist and increases with latitude.

The relatively constant elevation of the 0° isotherm implies a constant elevational transition from rain to snow throughout the year. In the Peruvian-Bolivian Andes, snowfall first occurs at the 3600-3800 m level. By 4300 m snow becomes as common as rain and above 4600 m rainfall is rare (Johnson, 1976). Because of the large diurnal temperature range snow melts quickly, except for accumulation on glaciers and perennial snow patches (Ribstein *et al.*, 1995), as afternoon air temperatures even at 4800 m temperatures are seldom below 0° C (Johnson, 1976). Within the model this transition is represented as follows: at elevations at or above the freezing level all precipitation is in solid form, at elevations above 5° C all precipitation is rain and a linear gradient in the fraction of precipitation falling as snow occurs at intermediate elevations.

The GEOS-1 climatology indicates the lapse rate varies considerably across the central Andes (**Figure 2.12**). In the humid eastern portions of the central Andes, the lapse rate is less than 6 °C km^{-1}, while in the arid central Andes it rises above 6.5 °C km^{-1}. Because convective activity determines the lapse rate in the tropics (Rennick, 1977), the lapse rate should be near the moist adiabtatic value in the humid eastern Andes. The GEOS-1 lapse rate in the humid eastern Andes is consistent with the observation that within 30 º of the equator the lapse rate are is within 20% of a moist adiabatic value (Stone and Carlson, 1979). A higher lapse rate in the western Andes is reasonable as the lapse rate in drier environments would be expected to deviate more from a moist value, although Webster and Streten (1978) point out that even here the lapse rate should be near a moist adiabatic value. General circulation model studies of Last Glacial Maximum conditions indicate that during full glacial conditions the lapse rate in the tropics was not significantly different than at present (Rind and Peteet, 1985).

Ablation in the mass balance model occurs at elevations where air temperatures are above freezing and is a function of the air temperature making representative air temperatures profiles for the humid and arid portions of the central Andes are required model inputs. The lapse rate determined from the GEOS-1 model are used to construct vertical two air temperature profiles in the mass balance model as follows. The highest elevation above the mean annual freezing level (z=0) where ablation occurs

Figure 2.12: Atmospheric lapse rates (°C km-1) between the 600 and 500 mb levels in the central Andes. The lapse rates are calculated from the GEOS-1 model for the years 1986-1990

corresponds to the elevation where the duration of melt (τ_m) is 0. Assuming a $\partial\tau_m/\partial z$ of -0.40 day m^{-1} (**Table 2.3**), this elevation is 450 m above the mean annual freezing level. In the mass balance model, as melting only occurs where air temperatures are above freezing, air temperatures at this elevation are 0° C as well. Air temperatures at elevations above and below are calculated using lapse rates of 6.0 and 6.5 °C km^{-1} for the humid and arid portions of the central Andes, respectively.

Sensible heat transfer is calculated using a bulk transfer approach. Sensible heat transfer is equal to the difference between the air (T_a) and surface (T_s, 0°C during melting) temperatures multiplied by a bulk transfer coefficient (α). Values of the transfer coefficient depend on surface roughness, air density, and wind velocity. The coefficient can vary greatly both spatially and temporally (see Kuhn (1979) for detailed treatments of turbulent exchange over a glacier). As the selected α has a large effect on the calculated mass balance profiles, two values of α, 1.7 MJ day^{-1} (~20 W m^{-2}) and 0.864 MJ day^{-1} (10 W m^{-2}) are used. The upper value is that suggested by Kuhn (1987) based on conditions in the Özetal Alps, Austria. The lower is that employed in glacial modeling of Oelermans and Hoogdorn

While there is some thermal seasonality in the central Andes, precipitation provides the primary seasonal climate signal in the region. The region is characterized by a wet-dry climate associated with the north-south migration of the Intertropical Convergence Zone (ITCZ). A single rainy

Table 2.3: Values used in parameterizing the mass balance model

Variable	Humid Andes	Dry Andes	Mid-latitudes	units
$\partial T_a/\partial z$	-0.006	-0.0065	-0.0065	°C m^{-1}
α	1.7 / 0.864	1.7 / 0.864	1.7 / 0.864	MJ day^{-1}
$\partial \tau_m/\partial z$	-0.40	-0.40	-0.10	day m^{-1}
Q_r (at 0° level)	0	0	0	MJ day^{-1} m^{-1}
$\partial Q_r/\partial z$	0.000	0.000	0.000	MJ day^{-1} m^{-1} m^{-2}
c (at 0°c level)	2000 / 1000	500 / 100	1000	mm
$\partial c/\partial z$	0.0	0.1	1 no rain-snow transition	mm m^{-1}

season extends from October to March and during this time most of the region receives between 70 and 90 percent of its annual total (**Figure 2.10**). Precipitation, even for the extreme westerly portions of the region, is derived from Atlantic moisture that has been transported across the South American lowlands. The regional precipitation pattern (**Figure 2.13**) shows the overall northeast-southwest decrease in precipitation from the Amazon lowlands (~2000 mm) to the Chilean coast (< 200 mm). Superimposed on this regional decrease are local orographic precipitation extremes occurring on lower Andean slopes. In response to this precipitation decrease, snowlines rise from east to west across the region.

The elevational pattern of precipitation in the tropics differs considerably from mid-latitude temperate climates and strongly affects the ELA response to a given climatic forcing. In most mid-latitudes mountain ranges, precipitation totals tend to increase with elevation to the highest peaks (Barry, 1981). However, in the central Andes a different pattern is evident. Rainfall tends to increase from the lowland plains to some elevation on the midslopes of the Andes with the zone of maximum precipitation at approximately 1000 m. Above this belt, the moisture content of the air diminishes and precipitation decreases with increasing elevation (Hastenrath, 1991).

Satoh (1979) noted that during the LLGM the maximum precipitation would be on the midslopes and not at higher elevations. If the vertical precipitation gradient along the eastern slopes is slightly negative, then

Figure 2.13: Mean annual precipitation in the central Andes (Hoffman, 1975). Contours are in millimeters. Shaded areas receive in excess of 3200 mm yr^{-1}

during the LLGM it is conceivable that more precipitation was received at the equilibrium line than at present. This occurs simply because as the ELAs of LGM glaciers moved lower, precipitation would continually increase.

In this model, the vertical gradient of precipitation in the eastern cordilleras is conservatively assumed constant with elevation ($\partial c/\partial z = 0$ mm m^{-1}). This is consistent with a slightly negative precipitation gradient observed at elevations 3500-4700 m in the eastern cordillera of Bolivia (Ribstein *et al.*, 1995). On the summit (6500 m) of Nevado Sajama, in the western cordillera of Bolivia, annual accumulations are approximately 500 mm (B. Francou, personal communication). Nearby at plateau level in Sajama village (4020 m) annual accumulation is approximately 300 mm (data from Servico Nacional de Meteorologia e Hidrologia, La Paz, Bolivia). Based on this decrease a slightly positive accumulation gradient (0.1 mm m^{-1}) was used for the drier western cordilleras. Compared to the mid-latitudes, these lower accumulation gradients increase the sensitivity of the ELA to temperature changes. Overall, imprecise knowledge of the exact value of the accumulation gradient ($\partial c/\partial z$) is a source of considerable uncertainty in this analysis. If the accumulation gradient is lower than is assumed (remembering the gradient can be negative as well, with precipitation decreasing with elevation), then the calculated temperature changes will be too large.

To a first approximation, elevational changes in incident long- and short-wave radiation can be neglected as the decrease in long-wave radiation with elevation is nearly offset by an increase in the short-wave component Kuhn (1989). In the model, radiation does not change with elevation. In reality the net radiation profile is strongly dependent on the vertical albedo profile. However, as noted by Kuhn (1989) when the ELA shifts so will the albedo profile; however, relative to the ELA, the albedo should remain approximately constant. The ELA response to climatic changes was found to be quite insensitive to the employed value of Q_r.

Another important glaciological difference between the central Andes and mid-latitude mountain belts is the coincidence of the accumulation (wet) and ablation seasons In the central Andes, the accumulation season coincides with the ablation season which results in a much stronger interdependence between accumulation and ablation processes in the tropics than the mid-latitudes. This interdependence is clearly demonstrated in the effects of ENSO on Bolivian glaciers. In the eastern cordilleras of Bolivia, ENSO years are characterized by lower precipitation resulting in decreased accumulation. However, the lower precipitation is accompanied by decreased cloud cover, and hence more incident solar radiation and higher temperatures which leads to increased ablation. The interdependence of ablation and accumulation make glaciers in the region quite sensitive to climatic changes during the wet/hot season, although changes in the dry season cannot be neglected (Francou *et al.*, 1995).

Modeling Results

Used in conjunction with the climatic inputs described (**Table 2.3**) the model reproduces the major features of the observed mass balance profiles in the central Andes (**Figure 2.14**). The steep mass balance gradient ($\partial b/\partial z$) of the ablation zone and the neutral balance gradient in the accumulation zone are well represented. The rise in ELA in response to decreasing precipitation is also captured in the model (**Table 2.4**). However, its magnitude is less than the observed east to west rise in the modern snowline. This discrepancy probably results from neglecting sublimation, which is an important ablation process in the arid central Andes, but is not understood well enough to be included in the model. As precipitation decreases, the sensitivity to precipitation change increases substantially (**Table 2.4**), but the temperature sensitivity remains high.

By altering the climatic inputs to typical mid-latitude alpine conditions (**Table 2.3**), the model is able to reproduce the observed differences between tropical and mid-latitude mass balance profiles (**Figure 2.14**). The ability to reproduce mass balance profiles in under variety of climatic settings suggests the model reasonably represents the glaciologically important aspects of alpine climate. The modeled response of a mid-latitude ELA to climatic changes (**Table 2.4**) compares well to other theoretical (Kuhn, 1981; Kuhn, 1989; Oerlemans, 1987) and statistical (Ohmura *et al.*,

1992) models of mid-latitude alpine glaciers. Glaciers in the eastern central Andes appear to have

Figure 2.14: Observed and synthetic mass balance profiles of **(A)** tropical and **(B)** mid-latitude glaciers. The synthetic mass balance profiles are calculated using the values in **Table 2.3** and an α of 1.7 MJ day^{-1}. Shaded areas indicate elevations at which melting occurs during a portion of the year. Sources for the observed mass balance profiles: Zongo (Francou *et al.*, 1995), Yanamarey (Hastenrath and Ames, 1995a), and Peyto and Hintereisferner (Kuhn, 1984).

a higher sensitivity to temperature changes and lower sensitivity to precipitation and net radiation changes than their mid-latitude counterparts.

Examining the pattern of modern snowlines (**Figure 2.7**) in light of these model results indicate the central Andes has essentially two glacial regimes that respond quite differently to climatic changes. In the humid eastern cordilleras the ELA is highly sensitive to temperature changes and relatively insensitive to net radiation and precipitation changes, while the western cordilleras are much more sensitive to changes in precipitation. The ability of the model to reproduce reasonable mass balance profiles under quite different climatic conditions lends confidence to the use of the model to address the two previously posed paleoclimatic questions: (1) how much colder and (2) how much wetter or drier was the central Andes during the LLGM?

Temperature Estimates

A large range of temperature and precipitation changes have been proposed for the tropical Andes on the basis of snowline depressions and other paleoclimatic proxy data (**Table 2.5**). Temperature change estimates for the late Pleistocene based solely on snowline depression range from as little as 3.5±1.6 °C based on glacier changes on the western slopes of the eastern cordillera Bolivia (Seltzer, 1992), to 10±1.9 °C based on cirque floor altitudes in Peru (Fox and Bloom, 1994), with numerous intermediate estimates. The large range results from (1) differences in the amount of

Table 2.5: Temperature and Precipitation change estimates for the central Andean region during the Last Glacial Maximum (LLGM)

Source	Method	Temperature Change (°C)	Precipitation Change	Location
Present Study	Snowlines	-5 to -9	wetter	Peru and Bolivia
Thompson *et al.* (1995)	$\delta^{18}0$ ice core records	-8 to -12		Huscarán, Cordillera Blanca, Peru
Stute *et al.* (1995)	noble gases in groundwater	-5.4±0.6		Piaui Provience, Brazil
Seltzer (1994)	Lacustrine sediments		wetter	Cordillera Quimza Cruz, Bolivia
Seltzer (1992)	Snowlines	-3.5±1.6		Cordillera Real, Bolivia
Fox and Bloom (1994)	Snowlines	-10±1.9	0-80% less (±25%)	Peru
Fox (1993)	Snowlines	-9.2±1.7	50-75% less (±25%)	Bolivia (18°-22°S)
Rodbell (1992)	Snowlines	> -5 to -6		Peru
Seltzer (1987)	Snowlines	-6.5 to -7	drier	Peru
Wright (1983)	Snowlines	-2	wetter	Junín Plain, Peru
Satoh (1979)	Snowlines	cooler	drier	tropical Andes
Hastenrath (1967, 1971)	Snowlines	cooler	wetter	tropical Andes

snowline depression used in the reconstructions, and (2) the methods used to convert an observed snowline change to a given temperature depression.

The smaller cooling estimates based on snowline depressions shown in **Table 2.5**, such as Seltzer (1992) and Wright (1983) are attributable to the very small snowline depressions, 300±100m and 300 m, respectively, used to infer a cooling. The larger cooling championed by Fox (1994) and Fox (1993) are attributable to two factors. The first is a slightly larger snowline depression (1400 m) than is assumed in this study. The second is the use of a non-linear temperature profile with a lapse rate that increases with elevation from approximately 6.5 °C km^{-1} at a height of 3.5 km to nearly 10 °C km^{-1} at 6 km. These two factors combine to produce the larger estimates of temperature depression. The remaining two studies with quantitative cooling estimates, Rodbell (1992) and Seltzer (1987) used a lapse rate of 6.5 °C km^{-1} and a snowline depression of approximately 1000 m.

The location in the central Andes where snowline would be expected to be most sensitive to temperature changes is along the humid eastern cordilleras. The consistent >1200 m snowline depression observed here is the best proxy for the temperature depression experienced in the region during the LLGM. Under humid conditions (**Table 2.4**), the modeled ELA lowering to a 1°C cooling is 167 m. Assuming the linear perturbation method can be extrapolated to larger snowline depressions, then the 1200 meter snowline depression can be explained by a 7.2 °C cooling. In contrast it would take a 10,000-20,000 mm increase in precipitation or a 200-400 W m^{-2} decrease in

net radiation to account for the depression, both of which are extremely unlikely. Because
the sensitivity of the ELA to precipitation and net radiation changes is so low, even if during the LLGM these quantities were different than at present, their effect on the estimated cooling would be small.

If the estimates of snowline depression are accurate to within ± 200 m, then there is a ±1.2° C uncertainty in the temperature estimates due solely to errors in estimating the snowline depression. Given this uncertainty, a cooling of 6.0-8.4° C is consistent with the mapped snowline depression. However, additional uncertainty in this temperature estimate arises because of lack of constraint of vertical climatic gradients and the bulk transfer coefficients. Considering these additional uncertainties, a reasonable conclusions is that snowline depressions in the central Andes indicate that the temperature during the LLGM was 5 to 9° C colder than present.

This temperature estimate is in good agreement with recent independent estimates of LGM temperature change including noble gas work of Stute *et al.* (1995) and evidence from treeline changes (van der Hammen, 1988). However, it is lower than the temperature calculated from oxygen isotopes in ice from Nevado Huscarán, Peru, Thompson *et al.* (1995). Nevertheless, modeled temperature depressions based on snowline evidence from the tropical Andes is consistent with the growing body of evidence that refutes the CLIMAP estimates of tropical SSTs.

Precipitation Estimates

Opinion is presently divided on the precipitation changes over the tropical Andes during the LLGM. Hastenrath (1967; 1971) and Wright (1983) argued that snowline depressions in the western cordillera were the result of increasing precipitation. The increase in precipitation was attributed to a southward shift in the ITCZ (Intertropical Convergence Zone) during the LLGM. Other workers (Fox and Bloom, 1994; Satoh, 1979; Seltzer, 1987) have argued the opposite: that precipitation in the western portions of the central Andes decreased during the LLGM.

Arguments for drier conditions in the central Andes during the LLGM hinge on the observation that the east-west snowline gradient was steeper during the LLGM than at present. The steeper gradient was attributed to the western cordilleras being relatively drier than the eastern cordilleras during the LLGM than at present. Three problems exist with this argument. The first is that measurements of the snowline gradient, especially the modern gradient, are imprecise. Second, over much of the Altiplano the amount of snowline lowering may be limited by the plateau topography and not by climate change.

Third, this approach assumes that the steeping of the snowline gradient is caused solely by increased aridity. This would only occur if the response of the snowline to a temperature change was uniform across the region. Previous glaciological work (Kaser, 1995; Kuhn, 1980) and the modeling presented in this study (**Table 2.5**) show this not to be the case. As the snowline rises in response to increasing aridity, it becomes less sensitive to temperature perturbations. Snowline in the western cordillera is simply less sensitive to temperature changes and more sensitive to precipitation change. This varying sensitivity would cause a non-uniform ELA lowering of the snowline across the central Andes without any change in precipitation. Steeper east-west snowline gradients during the LLGM could then be simply the result of a smaller shift in the ELA in the western cordilleras for a given temperature change. Snowlines changes do not indicate drier conditions in the central Andes during the LLGM. The extent of glaciers in the western cordillera today is precipitation limited. Having glacier expand extensively under much drier conditions (e.g. Fox, 1993; Fox and Bloom, 1994) is at odds with current understanding of modern glacier behavior in the region. Most authors conclude that glaciers in the arid portions of the central Andes are primarily sensitive to precipitation changes (Hastenrath, 1971; Kaser, 1995; Kuhn, 1980).

Because snowline elevation over the arid portions of the central Andes is presently limited by precipitation and not temperature, a precipitation increase is likely responsible for at least a portion of the observed LGM snowline depression. The upper limit of snowline in many portions of the central Andes is well the highest level where air temperatures are above freezing during the year (**Figure 2.15**). In the model the ELA will always lie below the highest position of the freezing level and thus cannot be used to

Figure 2.15: The highest elevation at which free air temperatures reach 0° C during the year in the central Andes. The elevations are calculated from GEOS-1 timeseries data for 1989. The contour interval is 100m. Shown in gray is the approximate area in which the lower limit of perennial snowcover exceeds the highest level of the 0°C isotherm.

estimate the ELA response under these conditions found in the western central Andes. Here sublimation is thought to be an important ablation process (Hastenrath and Kutzbach, 1985; Seltzer, 1993) on the cold dry glaciers as well as being responsible for the formation of spectacular penitents found here (Lliboutry, 1954). Because air temperatures never exceed the freezing level, to a first approximation, the ELA should have no dependence on air temperature changes. Under these conditions, a precipitation increase is the most plausible explanation for the 1000+ snowline depression. If precipitation is the major cause of snowline lowering in the western cordillera, the small area of high values (1000-1400 m) of snowline depression implies additional westerly precipitation along a small portion of the western Andes.

The existence of large paleolakes in the Altiplano basin at several times during the past has also been attributed to precipitation increases (Hastenrath and Kutzbach, 1985; Kessler, 1984). However, neither the lake nor glacial chronology is adequate to conclusively prove that the expansion of the two systems occurred simultaneously. Clayton and Clapperton (1995) concluded that a maximum glaciation and Paleolake Tauca, the youngest of the paleolakes were synchronous at 13,300 yr BP. However, even if lake and glacial expansion were synchronous, a large cooling (10-12 ℃) cannot be ruled out as an explanation for the existence of the lakes (Blodgett *et al.*, in press). Glacier and paleolake expansions are consistent with the hypothesis that wetter conditions existed on the Altiplano during the Late Pleistocene.

The modeling indicates that snowlines in the eastern and western cordillera of the tropical Andes respond to different climatic forcings. This underlines the possibility, as previously suggested by Hastenrath (1985), that snowline depression across the region could be asynchronous. However, lack of datable organic material in the arid regions severely limits the potential to test this hypothesis.

Summary and Conclusions

Comprehensive regional mapping of the modern and LLGM snowlines in the tropical Andes has revealed the magnitude and pattern of snowline depression in the region in a level of detail not previously possible. The general configuration of the regional snowline surface during the LLGM is similar to present with snowlines generally rising northeast-southwest in response to decreasing precipitation. This implies that atmospheric circulation was similar to present during the LLGM, with moisture coming from the east. In the eastern cordillera of the Peruvian-Bolivian Andes a consist 1200 meter snowline depression is observed. Snowline depression over portions of southeast Peru and Bolivia is much less, 500-800 m, and reflects non-climatic topographic controls on glacier expansion. The arid western cordilleras experienced a 1000+ m snowline depression similar to that in the east.

The sensitivity of tropical glaciers to climatic change is strongly dependent on their position relative to the 0 ℃ isotherm, which in the tropics is found at a constant level throughout the year. In the simple snowline depression model, this effect is captured in the model by the parameter τ_m, the duration of melting. In the eastern cordillera of the Andes, glacier ELAs are near the 0 ℃ isotherm level and experience melt during much of the year. The snowline elevation change in this region is sensitive to temperature changes. In the arid western cordilleras, the snowline rises considerably above the elevation of the 0 ℃ isotherm and is more sensitive to precipitation changes. Thus snowline response to a climatic change differs considerably across the region. The 1200 m snowline depression along the eastern slopes of the eastern cordillera is the best proxy for a temperature change experienced in the region and is best explained by a cooling of 5-9 ℃. The observed >800-1000 m snowline depression in the western cordillera in Bolivia cannot be explained by only a similar cooling, but also requires that precipitation increased in this area during the LLGM. These results together with several other converging lines of evidence imply a substantial cooling occurred in the South American tropics, which is inconsistent with the rather small (2 ℃) cooling indicated by CLIMAP sea surface temperatures.

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