CONTROLS ON STABLE OXYGEN ISOTOPE CONCENTRATIONS IN COROPUNA AND QUELCCAYA PERUVIAN ICE CORES OVER THE LAST 200 YEARS

A Thesis

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By

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ABSTRACT

Oxygen isotopes are useful proxy records in ice cores because of the selective fractionation process that occurs during evaporation and condensation of water molecules, yet the controls on these fractionation processes are under debate for tropical ice core records. Two ice cores from the eastern and western range of the Peruvian Andes (Quelccaya Summit Dome and Coropuna Caldera Core) are annually resolved for the last 200 years and provide an excellent means for comparison to localized instrumental meteorological records as well as regional measures of past climate. The oxygen isotope histories from these cores show no significant correlation with temperature or precipitation from two nearby meteorological stations or an automated weather station on the summit of Quelccaya. Yet significant correlation is found on a regional scale with Lake Titicaca water levels as well as equatorial Pacific sea surface temperatures over recent time. However, overall trends for the last century offer conflicting evidence to this end. On centennial and millennial time scales, temperature has been shown to be positively correlated with oxygen isotopes in tropical ice cores, yet the mechanisms for this control need further research.

Keywords: ice core; paleoclimatology; Quelccaya ice cap; Coropuna; oxygen isotopes; tropical glaciers; Peru; Andes
dedicated to my parents, family, and friends who have offered lifelong support and encouragement
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CHAPTER 1
INTRODUCTION

Tropical ice-core records provide valuable information about past climate, both on local and global scales. By preserving many indicators of past climate variability as well as specific circulation components (e.g. ENSO, Monsoons), tropical ice-core data help put past climate in perspective with current climate changes. The ability to forecast and model these changes is of great importance since tropical climatic variations on interannual and interdecadal time scales affect higher latitude climate (Baker et al., 2001). Not only does 70% of the world population reside in the tropics, but this region also plays a fundamental role in the global hydrological cycle (Hoffman 2003; Thompson and Davis, 2005). Therefore this climate record has important implications for those living in the tropics, as well as the rest of the world.

1.1 Andean Ice Core History

Since 1983 the Byrd Polar Ice Core group from The Ohio State University has recovered numerous tropical cores from around the globe. In South America, the OSU ice-core team has retrieved cores from four specifically selected areas located in the high elevations of the Peruvian and Bolivian Andes Mountains: Quelccaya (13°56'S; 70°50'W), Huascará (9°07'S; 77°37'W), Sajama (18°06'S, 68°53'W), and Coropuna
(15°32'S; 72°39'W) (Figure 1.1). Two cores in particular, the Quelccaya Summit Dome core (QSD) and the Coropuna Caldera Core (CCC) will be the focus of this thesis.

![Figure 1.1. Location of Andean ice cores produced by Byrd Polar Ice Core group (Thompson and Davis, 2005).](image)

Numerous ice core parameters are measured as proxies of past climate. Of particular interest are the isotopes of oxygen and deuterium which have shown strong relationships with temperature and moisture sources in Arctic cores (Dansgaard, 1964; Sharp, 2007). Oxygen isotopes are useful proxy records in ice cores because of the selective fractionation process that occurs during evaporation and condensation of water molecules. The relationship between oxygen isotope composition (δ¹⁸O) and temperature is well constrained for polar regions, but less straightforward for the tropics. One interpretation has been that oxygen isotopes (in the tropics) record precipitation
amount upstream (Broecker, 1997; Pierrehumbert, 1999). Yet it has also been shown that on longer time scales $\delta^{18}O$ in tropical ice cores is temperature dependent (Thompson et al., 2000; Thompson, 2001; Thompson et al., 2006). It seems possible that in the tropics isotope fractionation responds to both rainout (continentality effect) and temperature, but the temporal scale over which these factors play out is uncertain. Thus the relationship between $\delta^{18}O$ and temperature in the tropics is still under debate.

The purpose of this thesis is to gain a better understanding of the isotopic signature preserved in snow and ice from the tropics, including the extent to which regional and local climate factors play a role in these processes. This study includes the last 200 years of oxygen isotope records because both QSD and CCC are annually resolved for this period. Emphasis is placed on the last 100 years during which instrumental records have greatly helped improve our understanding of weather and climate. This provides a useful comparison for contemporaneous stable isotopic data retrieved from ice core analysis. By focusing on two ice core records located in relative close proximity to each other, the objective is to use instrumental meteorological records (both local and regional) to tease out temperature and precipitation controls on oxygen isotope records from tropical ice cores.

1.2 Ice Core Data Acquisition

In 2003 after collecting three new cores from the Quelccaya ice cap, an additional 3 cores were drilled on the new Coropuna site, a volcano southwest of Quelccaya. Cores 1 and 2 (~32 meters in length) were drilled to bedrock at an elevation of approximately 6450 masl, while Core 3 (~148 meters in length) was drilled to bedrock in the central
crater at approximately 6,410 masl. Each location was chosen for its simple flow regime to minimize alteration of the ice layers and produced multiple cores with annual resolution provided by distinct wet and dry season variability (Thompson et al., 1984, 1985, 1995; Thompson, 2000).

When Quelccaya was first drilled in 1983, logistics did not allow the cores to return to Byrd Polar Research Center frozen. At that time the samples were melted and bottled in the field, making them unsuitable for chemical analysis. Although a chemical analysis could not be completed, oxygen isotope concentrations from the bottled water samples were measured and provided a reliable history of the oxygen isotopic signature throughout the length of the core. Since the initial drilling program, a road was completed that facilitated transport of materials in and out of the field site, allowing cores from the 2003 drilling season to be delivered frozen to Byrd Polar Research Center and stored in the cold room facility (LGT pers. comm.). The samples were then cut and analyzed for stable oxygen and hydrogen isotopes ($\delta^{18}O$ and $\delta D$), insoluble dust concentrations and soluble aerosol chemistry. Detailed visual stratigraphy was also completed on each core. Annual resolution was obtained back to A.D. 470 in the QSD and back to A.D. 1760 in the CCC.

1.3 Data Sets

In addition to the isotopic compositions of QSD and CCC, seven other data sets are utilized throughout this thesis. Three sets of meteorological station data were analyzed; station Sicuani is located approximately 60 km to the southwest of Quelccaya, and Station Ccatcca is approximately 90 km to the northwest of the ice cap. A third, the
Quelccaya automated weather station, is located on the summit of the ice cap. Other data from the summit includes three years of snow pits and shallow cores. Two records of regional scale climate, Titicaca lake levels (a measure of central Andean precipitation) and the Oceanic Niño Index (a measure of equatorial Pacific sea surface temperatures) are also addressed.
The Andes Mountain chain in South America stretches 7000 kilometers along the west coast, comprising the longest chain of mountains in the world. With some peaks rising above 6700 meters in elevation and a mean elevation of approximately 4000 meters, this formidable barrier plays a large role in dictating South American climate. The cordillera (mountain chain) consists of an eastern and western range, the Cordillera Oriental and Cordillera Occidental respectively, separated by numerous intermontane basins throughout their length. The highest of these plateaus, known as the Altiplano, splits the cordilleras between 14° S and 22° S in the central Andes (Figure 2.1). The elevation of the Altiplano is on average 3800 meters, at approximately 650-600 hPa in the atmosphere (Johnson, 1976; Vuille et al., 2000; Garreaud, 2000).

To the west of the Altiplano, the mountains drop off quickly to the dry coastal plains which generally span only a few kilometers before meeting the ocean. Further north in areas of Ecuador and northern Peru the plain may reach 150 kilometers wide (Johnson, 1976). To the east, the Cordillera Oriental slopes more gradually to the continental lowlands. The Andes serve as a divide between two distinct climate regimes, that of the coastal plains and the continental lowlands.
2.1 Circulation and Seasonal Rainfall Patterns

The distinct climates occur because of the way in which the mountain chain intersects zonal flow aloft. The Andes north-south position perpendicularly intersects easterly tradewinds which entrain moisture laden air from the Atlantic. The barrier forces the air to rise and cool causing condensation and precipitation on the windward side of the mountain. As the air passes over the peak it has lost most of its moisture and also begins to subside and warm (Manabe and Broccoli, 1990; Garreaud et al., 2003; Aguado, 2004). By this process, the eastern mountain slopes receive moisture from the Amazon basin, while the west side remains fairly arid. Differences in moisture across the Altiplano are large, with rainfall averages more than 800 mm/yr in the northeast and less than 200 mm/yr in the southwest (Vuille et al., 2000; Garreaud et al., 2003)(Figure 2.2 a).
50-80% of the annual precipitation, deposited as snowfall in the upper Central Andes, occurs throughout the months of December through February (the Southern Hemisphere summer). Rainfall is mainly restricted to these months on the southwestern part of the Altiplano where DJF precipitation provides 75% of the annual total. In the northeast DJF rainfall accounts for 50% of the annual total, where the rainy season starts earlier and ends later, approximately November through April (Vuille et al., 2000)(Figure 2.2 b).

During the wet season, mornings are typically clear while the plateau is heated by solar radiation. By midday, this energy is enough to destabilize the air column and create buoyant air that will rise to the tropopause (the process of convection). In order for deep convection to occur, it is necessary that adequate amounts of water vapor exist in the atmospheric boundary layer (ABL), the layer of the troposphere which is in contact with
the surface of the earth. Observations have shown that when near surface water vapor exceeds \(\sim 5 \text{ g/kg}\), deep convection is possible over the Altiplano (Thompson et al., 1992; Hardy et al., 1998; Garreaud, 1999, et al., 2003).

Although enough radiation exists year round to destabilize the troposphere over the Altiplano (in the tropics radiation reaching the surface varies only slightly between winter and summer), the near surface water vapor threshold is rarely exceeded outside of the austral summer. Surface moisture over the Altiplano is ultimately controlled by upper level zonal flow patterns which influence slope flow on both sides of the Andes (Figure 2.3a)(Garreaud et al., 2003). Upslope flow occurs when a mountain side heats intensely during the day. The air over the slope warms and expands to higher altitude where it diverges (Aguado, 2004). Model results from Garreaud (1999) show that strong easterly flow at the 300-100 hPa level produce anomalously strong upslope flow (low level winds within the ABL) on the eastern flank, while limiting upslope flow on the western side. These low level winds bring moisture from the continental lowlands to the Altiplano, which is necessary for the formation of deep convection.

When upper level westerly winds are dominant, the opposite effect takes place, bringing dry air from the coast up to the Altiplano (Garreaud, 1999, et al., 2003; Lenters and Cook, 1999). During these times near surface moisture does not exceed the threshold and deep convection cannot take place. From May to October, westerly mid and upper level tropospheric winds dominate mean flow and limit rainfall on the Altiplano by keeping near surface moisture typically less than 2 g/kg (Garreaud et al., 2003).
Although a large moisture source, the Pacific Ocean, lies just west of the Andes, the southeast Pacific subtropical anticyclone and cool sea surface temperatures from the Humboldt current create dry and stable conditions. The cool moist air above the Pacific is restricted by a temperature inversion at approximately 800-900 meters above sea level which is sustained by subsidence throughout the year (Figure 2.3 b) (Garreaud et al., 2000, 2003; Vuille et al., 2000). Most moisture produced by the Pacific Ocean is driven up to the equator by the presence of the southeasterly tradewinds (Shimada et al., 1991). By these processes Pacific Ocean moisture does not reach the Altiplano and the Atlantic Ocean is thought to be the sole source for modern precipitation in this region (Garreaud et al., 2003).

Within the wet season, rain tends to be episodic, occurring in periods of about a week, separated by week long dry periods (Aceituno and Montecinos, 1993; Lenters and
Cook, 1999; Garreaud et al., 2003). This is thought to be reliant upon the availability of surface water vapor on the Altiplano rather than surface water vapor in the Amazon lowlands. When surface water vapor is below 3 g/kg dry episodes occur (Garreaud, 2000, et al., 2003). Associated with these dry and wet spells are changes in upper level circulation, particularly the strength and position of the Bolivian High. The upper-troposphere anticyclone centers around 15°S, 65°W during the rainy season and has been shown to migrate south and strengthen during precipitation episodes on the Altiplano. Conversely during dry periods the high pressure system moves northward and decreases in intensity (Lenters and Cook, 1999; Garreaud et al., 2003). It is not the movement of the Bolivian High that causes rainfall events, but the release of latent heat during condensation over the Amazon and Andes that influences the movement of the anticyclone (Lenters and Cook, 1999).

The duration of the rainy season is primarily dependent on the migration of the equatorial trough (a zone of convective cloudiness from low pressure that exists between the northeast and southeast tradewinds (Johnson, 1976; Aguado, 2004). As the equatorial trough migrates into subtropical latitudes, the subtropical jet stream weakens and moves southward (because of a weakened meridional temperature gradient), the Bolivian high becomes established (in response to convection) which all results in a southward expansion of the equatorial belt of easterly winds (Garreaud et al., 2003). The trough reaches its furthest position south in January, the time in which the Altiplano receives the most precipitation (Johnson, 1976; Shimada et al., 1991). The start of the rainy season
occurs progressively faster in its southward migration, yet moves westward more slowly, starting in September with thunderstorms occurring more frequently further into the wet season (Johnson, 1976).

Although the mechanisms producing precipitation are related to large scale circulation, the convective nature of the storms limits regions of precipitation, creating large variability across the Altiplano (Garreaud et al., 2003). For this reason rainfall amount can fluctuate greatly between weather stations in the area (Johnson, 1976).

2.2 Interannual Precipitation Variation

On an interannual basis as well as intraseasonal, large differences in precipitation amount occur on the plateau. Some rainy seasons can be extremely wet, while others can be extremely dry. This is a reflection of the number of wet or dry episodes within a season rather than a change in intensity (rainfall amount) brought by convective storms. Most of this variance is attributed to the El Niño Southern Oscillation (ENSO) phenomenon, which occasionally disrupts circulation over South America and beyond. In general, the negative phase of ENSO (El Niño) brings warmer temperatures and less precipitation to the Altiplano during the austral summer. The positive phase (La Niña) brings cooler temperatures and more precipitation (Garreaud and Aceituno, 2001; Garreaud et al., 2003).

Decreased snow accumulation has been recorded on Quelccaya Ice Cap (in southern Peru) during El Niño. Eight years of pit data from 1976 to 1983 record approximately 30 percent less precipitation than the annual average for the strong ENSO periods of 1976-1977 and 1982-1983. The below average precipitation events correlate
with anomalous warm sea-surface temperatures at Puerto Chicama, Peru (8°S, 80°W) (Thompson et al., 1984). Other measures of summertime precipitation, such as Lake Titicaca levels and rainfall station data also record drier El Niño years and wetter La Niña years. However, it is not uncommon for the opposite to occur and have a wet El Niño year or a dry La Niña year on the Altiplano (Garreaud and Aceituno, 2001; Garreaud et al., 2003).

The circulation patterns which bring about an anomalously wet or dry year are the same which create intraseasonal variation. Summers which register in the positive ENSO phase exhibit anomalous easterlies, as well as a strengthened and southerly displaced Bolivian high. Negative phase summers have anomalous westerly winds, reduced specific humidity, and a northerly displaced Bolivian high (Vuille, 1999).

Interestingly, atmospheric circulation in the tropical Andes is correlated with sea surface temperature anomalies in the tropical Pacific on interannual to interdecadal timescales (Bradley et al., 2003). The 200 hPa zonal winds above the Altiplano are part of an atmospheric response to ENSO patterns, dependent upon the size of SST anomalies and their spatial patterns in the Pacific Ocean. These upper level wind anomalies lead to precipitation changes across the plateau. Temperature anomalies over the Altiplano during ENSO events are also related to SSTA in the Pacific. During the El Niño phase, positive SST anomalies in the Pacific lead to greater evaporation and precipitation over the ocean. This increases the temperature of the atmosphere by the release of latent heat which transfers to near by regions such as the central Andes (Garreaud et al., 2003). Yet, the relationship between ENSO and precipitation over the Altiplano is not as straightforward. ENSO events vary greatly in the way sea surface temperature anomalies
develop and this strongly influences the location of the upper level wind anomalies, changing the way precipitation responds. As mentioned before, the Altiplano can experience wet El Niño years and dry La Niña years, opposite of what would be expected. Therefore, further work needs to be done to better comprehend the variability within the ENSO phenomenon and climatic responses to these events.
CHAPTER 3

STABLE OXYGEN AND HYDROGEN ISOTOPES

3.1 Oxygen and Hydrogen Isotopic Fractionation

Numerous ice core parameters provide proxy records of local and global climate. Stable isotopes, dust concentrations, aerosol chemistry and accumulation rates all help describe past conditions. Oxygen isotopes are of particular interest because of the selective fractionation process that occurs during evaporation and condensation of water molecules (H₂O). Oxygen has three naturally occurring isotopes (¹⁶O, ¹⁷O, ¹⁸O) while hydrogen has three naturally occurring isotopes (¹H, ²H, ³H), yet only two (¹H, ²H) are stable. Therefore water molecules can exist in 9 different forms, although water molecules with more than one heavy isotope are rare, and only 4 different isotopic combinations are frequent. The two molecules commonly studied in paleoclimatology are ¹H²H¹⁶O, also written as HDO, and ¹H²¹⁸O (Bradley, 1999; Sharp, 2007).

The isotopes of each element (oxygen or hydrogen) have similar chemical properties, but the varying mass allows for different bond strengths to other elements and different translational velocities. These distinctions drive the fractionation process between the various water phases. The vapor pressure of H₂¹⁶O is higher than HD¹⁶O and H₂¹⁸O, meaning the lighter molecules of water are more likely to break free from the water surface into the vapor phase. As water evaporates, the vapor becomes enriched in
$^{16}$O while the source water becomes relatively enriched in deuterium ($^2$H) and $^{18}$O. During condensation the opposite occurs. Cooling of the air mass drives molecules with lower vapor pressure (the heavier molecules), to preferentially condense first, leaving the vapor relatively depleted in $^{18}$O and the precipitate enriched (Sharp, 2007). While cooling continues, the vapor becomes increasingly depleted in heavy isotopes and so condensate is created with lower HDO and H$_2$$^{18}$O concentrations than when the process started. This provides a direct relationship between air temperature and isotopic concentration (Bradley, 1999).

Evaporation is a kinetic process and the fractionation factors depend on numerous variables including local conditions of humidity, turbidity of the liquid, and diffusion rates of hydrogen and oxygen from further within the liquid (References within Sharp, 2007: Craig and Gordon, 1965). At equilibrium, vapor above the ocean would contain 10‰ less $^{18}$O than average ocean water, yet in actuality the vapor is 3-4‰ more negative than this, giving vapor above the ocean $\delta^{18}$O values of -13 to -11‰. This disparity exists because evaporation is not an equilibrium process; water above the oceans does not reach 100% saturation, so therefore evaporation is unidirectional, making the vapor lighter than predicted. On the other hand, condensation is an equilibrium process, and the fractionation factor depends on temperature alone (Sharp, 2007).

Isotopic ratios are traditionally written as the ratio of the heavy isotope to the light isotope, in this case $^{18}$O/$^{16}$O. SMOW (Standard Mean Ocean Water) is used as a reference for oxygen isotope analysis and the isotope composition is written as a deviation ($\delta$) from this value, in per mil units (‰). The deviation is defined as: $\delta^{18}$O =
\[
\left[ \frac{(^{18}\text{O} / {^{16}\text{O}})_{\text{sample}} - (^{18}\text{O} / {^{16}\text{O}})_{\text{SMOW}}}{(^{18}\text{O} / {^{16}\text{O}})_{\text{SMOW}}} \right] \times 10^3 \%
\]
Measurements of the \(^{18}\text{O} / {^{16}\text{O}}\) ratio are made by a mass spectrometer and error is usually limited within \(+/-\) 0.1\%. (Bradley, 1999).

Because fractionation of oxygen and deuterium usually occur in step, there is a very near linear relationship between the two in average values of meteoric water (liquid or solid water that has fallen from the sky). This is known as the global meteoric water line (GMWL) (Figure 3.1), and is represented by \(\delta D = 8\delta^{18}\text{O} + 10\) in modern day (Craig, 1961; Bradley, 1999; Sharp, 2007).

![Figure 3.1 Deuterium and oxygen-18 in meteoric waters, plotted as their deviation from standard mean ocean water (SMOW), units are per mil (‰) (Craig, 1961).](image)

Figure 3.1 Deuterium and oxygen-18 in meteoric waters, plotted as their deviation from standard mean ocean water (SMOW), units are per mil (‰) (Craig, 1961).
3.2 $\delta^{18}$O-Temperature Relationship

Dansgaard (1964) first observed the strong relationship between $\delta^{18}$O and surface temperature in the polar regions, where precipitation forms near the land surface. When conditions are colder, $\delta^{18}$O values are depleted, and when conditions are warmer, $\delta^{18}$O becomes enriched, creating a positive correlation between $\delta^{18}$O and temperature. The relationship can be defined as $\delta = a T_s + b$ for a specific region. Dansgaard (1964), in regards to polar areas, assigned a slope (a) of 0.67‰ per °C. It was originally thought this relationship held over time, yet studies by Cuffey et al. (1995), show the slope may be different for the last few millennia, than it was for older ice (Thompson, 2000). Over recent time, numerous analyses of meteoric water from all over the world have produced a relationship of $\delta^{18}$O = 0.69$T_{\text{average}}$ – 13.6 (Sharp, 2007).

There is still much debate as to whether a $\delta^{18}$O-temperature relationship holds at low latitudes (such as Coropuna and Quelccaya). Here, precipitation forms in large convective systems (compared to near surface level formation in the polar regions) and seasonal temperature ranges are relatively small. Yet, it has been shown that on century and longer time scales isotopic variability positively correlates with temperature (Thompson, 2000; Thompson et al., 2006). The presence of the Little Ice Age and Medieval Warm Period in the Quelccaya $\delta^{18}$O record provides notable evidence of this relationship (Thompson et al., 1986; Thompson, 2000; Thompson and Davis, 2005).

On seasonal time scales temperature changes are minimal in the tropics. The range on the Quelccaya summit is 3.76°C (Quelccaya automated weather station) and only about 2°C for sea level temperatures (Thompson et al., 1984), still a large seasonal
$\delta^{18}O$ cycle is evident. The $^{18}O$ composition of Andean ice cores shows depletion in the austral summer (wet season) and enrichment during the winter (dry season), a relationship opposite of that observed in high latitudes. In many cases, isotopic composition of precipitation may vary throughout seasons due to differing source waters. Yet, it has been shown that the tropical Atlantic Ocean is the unique source of precipitation for the Central Andes year round (Grootes et al., 1989).

3.3 Models of Isotopic Fractionation across the Amazon Basin and Central Andes

Two models have thus far been presented to explain the seasonal variance of $\delta^{18}O$ in tropical Andean ice cores (Thompson, 2001). Grootes et al. (1989) used a hydrological balance model with Rayleigh fractionation to recreate moisture transport as it moves away from the Atlantic Ocean, over the Amazon basin and up the Andean slopes to the Quelccaya ice cap (Figure 3.2). Rayleigh fractionation assumes condensation is an equilibrium process and that all condensate is directly removed from the system, a simplification of what actually occurs in convective cloud systems.

Atlantic Ocean water starts out with an average $\delta^{18}O$ value of $+1\%_o$ (Grootes et al., 1989). As this water evaporates, the lighter isotopes are preferentially concentrated in the vapor. The air mass makes its way across the Amazon basin and when rainout occurs the heavier isotopes condense first and fall in precipitation. The more rainout that occurs, the more depleted in $^{18}O$ the vapor becomes, which in turn creates precipitation more depleted in $^{18}O$. This process of continued rainout and $^{18}O$ depletion over land is known as the “continentality effect.”
Using the Rayleigh equation: \( \frac{R}{R_0} = F^{(\alpha - 1)} \) (where \( R \) and \( R_0 \) represent the current and the original isotopic abundance ratio of the water vapor, respectively, and \( F \) represents the fraction of the original amount of water vapor remaining in the air), while keeping \( \alpha \) (the isotopic fractionation factor) constant, it is possible to calculate \( \delta^{18}O \) depletion that occurs from water vapor loss of the air mass in its trajectory over the Amazon basin. From this, Grootes et al. predicted an isotopic depletion of 18.2-15.5‰ during the wet season, and nearly no depletion during the dry season. By the time the vapor reaches the base of the Andes, its mean \( \delta^{18}O \) value is -20‰. As the air mass rises over the slopes, adiabatic cooling forces further condensation and further depletion of \( ^{18}O \) (known as the altitude effect). Fractionation rates caused by the altitude effect remain similar for both the wet and dry seasons. In the 5000 meter rise from the Amazon basin...
to the Quelccaya ice cap, an additional 10‰ depletion occurs (Grootes et al., 1989). Using Grootes’ estimation of 0.2‰/100m depletion up the Andean slopes, precipitation falling on Coropuna would be further depleted by 1.52‰ when only considering the altitude effect (although further continentality effect would take place as well). Yet as discussed in Chapter 4, a lapse rate of 0.2‰/100m may not be a reasonable estimate.

Post-depositional processes may also affect the isotopic composition measured in tropical ice cores. During the dry season enrichment at the surface of the snow may increase the seasonal cycle shown by δ¹⁸O, while meltwater can smooth the isotopic signal (causing a dampening of the high and low isotopic concentrations). The effects of continentality, altitude and postdepositional processes can be considered additive. Grootes et al. found their predicted values of ¹⁸O depletion (-4.1 to -33.7‰) to be in agreement with observed values recovered from snowpits on Quelccaya (-8.0 to -30.8‰). Thus, using a hydrological balance model, Grootes et al. (1989) accounts for ¹⁸O depletion on Quelccaya by rainout upstream.

Vuille et al. (2003a,b) notes that although the overall mechanisms in this model are correct, newer evidence necessitates readdressing some of the assumptions. In regards to the dry season, it was found that depletion of air masses moving over the Amazon basin occurs during both the wet and dry seasons, albeit with a much lower continental gradient in the dry season. More recent studies of postdepositional processes on stable isotopic composition on the surface of tropical glaciers show that most of the enriched layer created during the dry season is removed from the surface by sublimation (Stichler et al., 2001; Vuille et al., 2003b; Hardy et al., 2003).
The second model to explain seasonal variance of $\delta^{18}$O in tropical Andean ice cores, accounts for the different mean condensation levels during the different seasons. In the tropics, precipitation arrives in the form of large convective systems, with much greater vertical ascent than cloud formation in the polar regions. In the wet season, the condensation level is approximately 2 kilometers higher than during the dry season. When the mean condensation level is higher, condensate forms at cooler temperatures and therefore the isotope values should be more depleted in the wet season. In the dry season, with condensate forming in lower portions of the atmosphere, temperatures are warmer and condensate is relatively enriched (compared to the dry season). The pattern in isotopic concentrations found on Quelccaya could occur by this process (Thompson et al., 2000). This model shows the possibility that atmospheric temperatures drive isotopic composition in tropical glacier records.

It is possible that some combination of these two models explains the isotopic composition of precipitation, and even that precipitation amount and temperature are not mutually exclusive. On interannual timescales, Andean climate has been warm and dry or cold and wet due to the influence of the tropical Pacific Ocean. During times of prolonged cold SSTA, upper level wind anomalies over the Andes create wet conditions, while periods of prolonged warm SSTA create dry Andean conditions (Vuille et al., 2003b; Garreaud et al., 2003).

Yet this relationship is not completely straightforward as some tropical glaciers may also be recording local conditions superimposed on regional conditions. Glaciers located in relative proximity to one another, such as Huascarán and Sajama, have shown distinct climate differences in the past. While Huascarán records a warm and wet
Holocene, Sajama shows a warm and dry Holocene. Also, it is likely that Sajama recorded the previous existence of large lakes upwind from the mountain, which could have contributed to an increase in precipitation during LGM and also into the early Holocene until the lakes dried out (Thompson and Davis, 2005). Therefore understanding the parameters influencing $\delta^{18}O$ for each particular glacier are important when comparing Andean glaciers, as well as discussing interpretations of past regional scale circulation.

The negative relationship between $\delta^{18}O$ and temperature found on a seasonal timescales, and the positive relationship between $\delta^{18}O$ and temperature found on decadal and longer timescales, renders the question, at what timeframe does temperature become an overriding factor and why does this occur? This thesis will address this question by comparing short term meteorological records to Andean glacier isotopes in Chapter 4, and then will discuss these results within the perspective of longer term isotopic records in Chapter 5.
CHAPTER 4
RESULTS AND DISCUSSION

4.1 Quelccaya and Coropuna Isotope Record 1800-2003

Over the last 200 years, relative oxygen isotope levels in the Quelccaya Summit Dome core and the Coropuna Caldera Core are annually resolved. Out of the ten ice core drilling sites between 0° to 52°S, only Quelccaya, Coropuna and Huascarán are annually resolved until at least 1800 with low uncertainty (Vimeux et al., 2008; Mary Davis pers. comm.), making the comparison between Quelccaya and Coropuna a unique study.

In the following discussion of ice core records it is imperative to understand that the information retrieved is mainly a representation of conditions during precipitation events. For tropical ice cores this translates to a representation of conditions mostly during the wet season, rather than annual mean conditions. Because the wet season is of greatest interest, the unit of “thermal year” will be designated in discussions of isotopic records and comparisons with concurrent meteorological data. A thermal year will be defined as July 1 to June 30 of the following year. This measurement delineates years by the height of the dry season layer and includes the length of a full wet season. Thermal years will be noted as the later year, for example, thermal year 2005 runs from July 1,
2004 to June 30, 2005. In comparisons with weather station data, the concurrent wet season (November-April) is used. As with thermal years, wet seasons will be classified as the later year.

Despite Quelccaya and Coropuna being approximately 350 kilometers apart and separated by 760 meters in elevation, their oxygen isotope records show excellent similarity over this time (Figure 4.1). From 1800-2003, the average $\delta^{18}O$ level at Quelccaya is $-17.97\%$, while Coropuna is $-18.35\%$. This is only a $0.38\%$ difference for the 200 year period.

![Figure 4.1 $\delta^{18}O$ values in QSD and CCC from 1800-2003.](image)

Not only do these cores have strikingly similar oxygen isotope averages for the last 200 years, but they both show a similar enrichment during the most recent century.
From 1800-1900, Quelccaya and Coropuna’s isotopic averages are -18.46‰ and -18.88‰, respectively. From 1900-2003 Quelccaya and Coropuna’s the averages are -17.48‰ and -17.80‰, both showing an enrichment of about 1‰.

The significance of the similarity of the records and shared enrichment is twofold. One, this demonstrates that the two glaciers are recording similar climatic events within their ice, and these events are more predominant in controlling δ¹⁸O than local scale climate events. Two, the isotopic signature as recorded by Quelccaya and Coropuna should theoretically be offset from each other by the fractionation processes of continentality and altitude. Yet no offset is apparent in the oxygen isotope records of these two cores. Over the last 200 years the two cores record extremely similar δ¹⁸O levels. Some other process or combination of processes is taking place that negates the fractionation processes of continentality and altitude, which would otherwise be apparent in the δ¹⁸O signal. The paragraph below will outline the theoretical fractionation that should occur between the two sites, followed by a discussion about the implications of the similar record between these sites.

Because Coropuna lies approximately 350 km to the southwest and 760 meters in elevation higher than Quelccaya, the isotopes should record a fixed depletion that occurs from the effects of continentality and altitude, given that prevailing winds from the northeast move moisture first over Quelccaya, then Coropuna. Poage and Chamberlain (2001) determined that there is a near linear relationship between the depletion of δ¹⁸O and increasing elevation. Most regions of the world adhere to a -0.28‰/100m, while Central and South America average a slightly higher lapse rate of -0.30‰/100m. Yet this relationship is much better constrained for elevations less than ~5000 m. They note there
is much more scatter in the high altitude groupings and hypothesize the variation could be due to post-depositional changes in snow or even the influence of a secondary moisture source from the upper troposphere. Nonetheless, their lapse rate for the high altitude group was determined to be -0.41‰/100m. However in their compilation of studies considered in the paper, lapse rates at the upper ends of -0.77‰/100 m were noted in the Cordillera Blanca region of Peru. This leaves a possible range of lapse rates between -0.41‰/100m to -0.77‰/100m for the area of interest. Using both end members as possible lapse rates between Quelccaya and Coropuna yields an isotopic depletion between 3.20‰ to 6.01‰ for the 760 meter difference between the two sites. Even using the lowest possible lapse rate for lowland areas (-0.28‰/100m) there should still be a depletion of 2.18‰ between the two sites. Yet these two records are only averaging a 0.38‰ difference over the 200 year record. In fact, Coropuna’s isotopic record is more depleted than Quelccaya for only 113 years out of the 204 years on record, meaning 45% of the time Coropuna’s isotopic values are actually more enriched than Quelccaya (Figure 4.2).

It is apparent other fractionation processes are controlling the isotopic signature in the record and offsetting the difference of continentality and altitude that should consistently create a more depleted record in Coropuna ice. A few possibilities exist for explaining this phenomenon. One, because Coropuna is a much drier site than Quelccaya, the post depositional process of sublimation could play a larger role. If sublimation is much greater at this site, then enrichment of the isotopic signature would
be enhanced. Therefore, although the snow falling on Coropuna may actually be further depleted as would be expected through fractionation processes, sublimation enriches the end isotopic result creating similar values to that of Quelccaya.

![Graph showing difference between CCC δ¹⁸O and QSD δ¹⁸O](image)

**Figure 4.2** QSD annual isotopic values subtracted from CCC annual isotopic values. Areas above the zero line indicate thermal years when Coropuna was more enriched than Quelccaya, while negative values indicate years when Coropuna was more depleted than Quelccaya.

Stichler et al. (2001) conducted a sublimation study on Cerro Tapado, another of the cold high-altitude glaciers in the tropical Andes. They note that sublimation only affected layers down to a depth of 5-10 cm, even under extreme dry environmental conditions. More importantly, they suggest total mass loss at the surface must be taken into account since this eliminates the entire ice crystal, therefore erasing the altered part of the record. Because of this, the effects of sublimation on the isotope signature in an ice core are limited. If however, an isotopic record has been altered by the effects of sublimation, it should be identifiable by comparing concurrent measurements of δ¹⁸O and
Post-depositional changes in δD occur by differences in diffusivity of δ¹⁸O and δD through the firn layer. In plotting the δ¹⁸O-δD data, sublimation is detectable if the slope is less than 8 (that of the meteoric water line) (Stichler et al., 2001).

Neither Coropuna nor Quelccaya δ¹⁸O-δD plots (Figure 4.3) have slopes of less than 8. In the Cerro Tapado experiment, data which had a slope of 4.88 defined a typical sublimation line. All Coropuna and Quelccaya data fall on a very tight line with a slope close to 8, indicating that sublimation has not affected the annual averages of the ice core isotopic data. By analyzing the δ¹⁸O-δD plots for indications of the influence of sublimation on the isotopes, it does not seem likely that the post-depositional process of sublimation affects the final isotopic value extracted from the ice cores.

Secondly, it is possible that the central Atlantic (and Amazon Basin) is not the only true moisture source for precipitation on Coropuna as previously thought. Coropuna’s location on the summit of the Western Andean Cordillera leaves open the
chance that precipitation from the Pacific makes its way up to the glacier. If this were the case, precipitation from the Pacific would be enriched relative to Atlantic moisture because it has not experienced rainout across the scope of the continent. Mixing in a fresh precipitation source into the Atlantic vapor source would serve to enrich the record preserved in the snow, the end result being isotopic values more similar to those of Quelccaya.

Whatever process may negate the theoretical offset in oxygen isotopes between the two sites, it is a process that has remained constant over the last 200 years. The true importance lies in the similarity of the two records and that the glaciers are recording similar climatic events. The similarity of the isotope record between the two sites confirms the isotopes within the glacier ice are recording something larger than local scale climate events. But over just how large of an area do the isotopes integrate the climate signal and what type of climate events are being recorded?

In order to answer these questions, the rest of this chapter focuses on comparisons between δ¹⁸O and South American meteorology records of precipitation and temperature as well as other proxy data. These comparisons are evaluated on both regional and local scales to assess the most important forcing on oxygen isotopes as recorded by tropical glaciers.

**4.2 Local Scale Records of Precipitation and Temperature**

**4.2.1 Quelccaya Automated Weather Station**

Since 2003 an automated weather station has been in operation on the summit of Quelccaya as a joint project between the University of Massachusetts and The Ohio State
University. This time series provides a unique look into the environment on the ice cap during periods when weather conditions make it almost impossible for direct observance. The information obtained during the wet season also provides essential data to help interpret what portion of annual conditions is recorded by the ice core.

The solar powered station consists of a broad array of sensors recording data which are transmitted by satellite. Although the station is automated, necessary maintenance has been performed three times, each time raising the station’s sensors high enough to account for the next season of accumulation (pers. comm. Doug Hardy) (Figure 4.4).

Figure 4.4 Maintenance of the Quelccaya AWS includes raising the sensors each dry season to account for accumulation in the wet season. This photo shows work completed in the 2007 dry season.
Temperature

From 7/22/04 until present, temperature has been recorded hourly, while accumulation (recorded as height change from a datum) is recorded daily. The average temperature on the summit over this time series (7/22/04 – 3/22/08) was -4.47°C, with an average daily range of 6.20°C and an average seasonal range of 3.76°C. Typical for mountainous regions in the low latitudes, the temperature difference between day and night is much greater than the difference between seasons. Figure 4.5 displays monthly mean values of air temperature. The warmest months on Quelccaya’s summit were generally January and February (average -3.0°C) during the height of the wet season, while the coolest months occurred during June and July (average -6.78°C) in the dry season.

Thermal Year Average Temperatures

2005: -5.06°C
High Feb 05: -3.30
Low Jun 05: -8.00
Range: 4.70

2006: -4.51°C
High Feb & Oct 06: -3.40
Low Jul 06: -6.04
Range: 2.64

2007: -3.86°C
High Jan 07: -2.31
Low July 07: -6.27
Range: 3.96

Figure 4.5 Monthly mean air temperature on the summit as recorded by the Quelccaya automated weather station. Annual averages, seasonal highs, lows and range are listed in the left column.
Most likely the higher than average temperatures of thermal year 2007, shown in Figure 4.5 occur because of El Niño conditions which began early 2006 and lasted until early 2007. Slightly lower temperatures in the wet season of 2008 probably stem from weak La Niña conditions developing at the time, which created Pacific equatorial sea surface temperature anomalies up to -1.5 (normalized value) during December, January and February. These temperature trends with ENSO are typical for the Altiplano region and further substantiate this relationship on the summit of Quelccaya. Thermal years 2005 and 2006 may be more typical of a “normal” year on the summit because ENSO conditions were more neutral during this time than in 2007 and 2008. Note that average temperatures recorded in February 2005 through June 2005 should be interpreted with care due to problems with minimum temperatures as read by the automated weather station, although it is most likely the period affected is actually limited to the month of February. Over this time the snow surface rose high enough for the sensors to be in the boundary layer (within ~ 0.5 meters of the snow surface) where near surface inversions may affect temperatures (Doug Hardy; Ray Bradley, pers. comm.). Yet the average readings over the possible problematic period seem in reasonable agreement with nearby weather stations (see section 4.2.2), and therefore may not be an issue.

Although average temperatures stay below freezing, days with numerous hours above freezing now occur on the summit. A suite of shallow cores drilled in 1979 indicate that at this time the 0º isotherm was somewhere just below 5600 meters, below the ice cap summit. But by 1991 the 0º isotherm had risen approximately 100 m over the
12 year period, as indicated by smoothing in the $\delta^{18}O$ signal from short cores drilled on the summit in 1991 and 1995 (Figure 4.6). This smoothing of the isotopic signal signifies that percolation of melt water has occurred (Thompson et al., 2000).

![Figure 4.6 A suite of shallow cores drilled in 1979, 1991, and 1995 indicating the smoothing which has begun to occur from percolation of meltwater on the glacier (Thompson et al., 2000).](image)

In a study of radiosonde measurements in the tropics (~15° N - 15°S) and numerical model simulations, Diaz and Graham (1996) found an increase in freezing level height from 1970 to the late 1980’s of about 110 m, quite similar to the results of Thompson et al. (2000). They also found that some of the greatest simulated warming occurred in the area of Peru where Quelccaya is located. The rise of freezing level heights is consistent with the documented melting now occurring on Quelccaya and most
dramatically apparent in the retreat of its outlet glacier, Qori Kalis (Thompson et al., 2000, 2006). In conjunction with this previous data, the AWS hourly temperature profile further documents the above freezing temperatures which are now commonplace.

Figure 4.7 Number of hours above freezing for each day during the period of observation for the Quelccaya AWS.

Figure 4.7 shows the number of hours above freezing that occurs on a particular day. It is now a common occurrence during the wet season to have days with 4-7 hours above freezing. In the 2005 wet season, 24 days had 4-7 hours above freezing, while the 2006 wet season had 7 days, and the 2007 wet season had 32 days. Even during the colder dry season, numerous days reach above 0°C. On average 5 days out of each wet season reached above freezing, although May 2007 had 10 days reach above 0°C. 2007 had a large spike in hours above freezing, with 393 hours total occurring over a total of 151 days, once again most likely due to effects of El Niño. Yet 2005 and 2006, which can be considered more neutral years still had 248 hours and 146 hours above freezing.
respectively. The highest temperature recorded over this time period was 2.8°C on November 10, 2007, although the mean temperature for all hourly readings above 0°C (a total of 872 hours) was 0.65°C.

The effect which melting and percolation have on the overall isotopic average of the ice core record is an important area of further research. A comparison of the 1983 Quelccaya core record with the 2003 Quelccaya core record (drilled after melting had begun on the summit) shows excellent reproducibility (Figure 4.8) and ensures that not all data quality is affected by the melting.

![Figure 4.8 A comparison of the Quelccaya 1983 core δ¹⁸O and the Quelccaya 2003 core δ¹⁸O (drilled after melting had begun on the summit) (Thompson et al., 2006).](image)

Although melting conditions on the surface began around 1991, the isotopes further down core were not influenced (Thompson et al., 2006). The main concern now is smoothing of the isotopic values in the upper levels which are affected by meltwater percolation. This issue is briefly addressed in section 4.2.3, but further years of shallow cores on the summit would be helpful in understanding the process.
Precipitation

Because there are numerous difficulties of measuring solid precipitation, especially in remote areas, changes in snow surface height from one measurement to the next were taken as a proxy for accumulation or ablation on the summit. This procedure was used and published in Hardy et al. (2003) for the Sajama summit, and this paper will make the same assumptions. The height profile obtained from the automated weather station results from a range of accumulation and ablation processes. Surface height increase can be caused by snowfall and/or wind redistribution of snow, while surface height decrease may occur from wind scour, settling, melting and sublimation (Hardy et al., 2003). It is important to understand the interactions of these processes, for their combination is what produces a net accumulation stratigraphy. During the wet season on Sajama precipitation overwhelms ablation processes and only a few days of surface decrease occur between precipitation events. As the wet season comes to a close, surface-lowering processes take over. As the snow pack ages over time (during the dry season), decreased rates of settling occur, scour becomes less likely to remove mass, and decreasing temperatures and vapor pressure reduce the chance of melting (Hardy et al., 2003). Processes would be similar on Quelccaya with a few exceptions. The wet season is longer (October through April) and so dry season processes would occur over a much shorter season. Also, scour is probably not as much of an issue on Quelccaya because layering is found to be uniform throughout the yearly layers. With scour being less of a factor to remove snow, decreasing temperatures and vapor pressure would still favor sublimation as the main dry season process which lowers the snow surface.
Figure 4.9 illustrates the periods of intense accumulation (generally October through April on the Quelccaya summit) followed by periods of mass loss initially from settling, and increasingly from sublimation further into the dry season (May through September). Figure 4.9 also highlights the importance of the dry season lowering processes, that is, snow which falls late in the wet season is generally ablated, and therefore not represented in an ice core record (Hardy et al., 2003).

![Daily snow surface height at Quelccaya AWS](image)

**Figure 4.9 Snow surface height record at Quelccaya AWS relative to height of datum.**

Monthly accumulation during the wet season varies greatly year to year, ranging from a few centimeters to over 50 cm. During the dry season, observations show a range of less than a cm to over 20 cm of surface lowering per month (Figure 4.9). Thermal year net accumulation is as follows, 2005: 1.97 meters, 2006: 2.06 meters and 2007: 2.08 meters. Considering total snowfall (summing months with only positive accumulation) did not change precipitation amounts greatly (2005: 2.04 m, 2006: 2.08 m, 2007: 2.14 m), although these numbers are a bit misleading. Figure 4.10 demonstrates the importance of...
including total dry season loss into the previous season’s accumulation data. An example is given below to show why the delineation of thermal years is not appropriate in this case.

<table>
<thead>
<tr>
<th>Month</th>
<th>Accumulation (meters of snow)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1/1/05</td>
<td>-0.4</td>
</tr>
<tr>
<td>7/1/05</td>
<td>-0.2</td>
</tr>
<tr>
<td>1/1/06</td>
<td>0.0</td>
</tr>
<tr>
<td>7/1/06</td>
<td>0.2</td>
</tr>
<tr>
<td>1/1/07</td>
<td>0.4</td>
</tr>
</tbody>
</table>

2005: 2.04 m  
2006: 2.06 m  
2007: 2.14 m

Figure 4.10 Net monthly accumulation or surface loss as recorded by the Quelccaya AWS. Right column compares total accumulation to net accumulation for years 2005-2007.

The winter months of 2005 (May, June, July, August and September) recorded much more surface loss than either of the dry seasons following it, therefore decreasing the net accumulation in 2005 by almost half a meter. After July 1 (considered the start of the next thermal year) a period of approximately 3 months continues in the ablation of the surface snow, further decreasing the net accumulation. Accumulation as recorded in the ice core is more likely the process occurring from sometime around October until August or September. For example, it is more likely that the ice core would show net accumulation for these three years as follows, 2005: 1.57 meters, 2006: 1.97 meters and
2007: 1.86 meters although total accumulation was actually very similar for all three years. These results imply that care must be taken when using net accumulation as a measure of precipitation amount on the snow cap. Bradley et al. (2003) noted that the amount of accumulation which is left at the end of each dry season can vary from year to year, but in all years, most of the late summer season snowfall is not retained in studies on the Sajama glacier. Even though Quelccaya has a longer wet season and receives more precipitation than Sajama, Figure 4.10 shows that some end of season snow is not retained on Quelccaya as well.

Also of interest is 2007 total accumulation in which El Niño conditions existed, yet accumulation was not affected. Precipitation at two nearby weather stations, Sicuani and Ccatcca, was also near average. Although it is not common, ENSO events can be anomalous, such as the El Niño of 1972/73 which brought warm but wet conditions to the central Andes, and the La Niña of 1988/89 which was a dry year (Garreaud and Aceituno, 2001).

### 4.2.2 Sicuani and Ccatcca Weather Stations

The next closest meteorological stations to Quelccaya with the longest temperature and precipitation records are Sicuani (14°14'14.2" S, 71°14'12.1" W, 3546 m asl) and Ccatcca (13°36’35”S, 71°33’36”W, 3729 m asl). Station Sicuani is located approximately 60 km to the southwest of the Quelccaya ice cap and Station Ccatcca is approximately 90 km to the northwest. Total monthly precipitation (mm), and average monthly temperature (°C), have been reported since the mid 60’s, although the records are not complete throughout (Figure 4.11 a, b, c, d).
Figure 4.11 Wet season precipitation (November-April) as measured at (a) Station Sicuani (b) Station Ccatcca, wet season average temperature (November-April) as measured at (c) Station Sicuani (d) Station Ccatcca. Red line indicates linear trend over time period.

Over the past 40 years wet season precipitation and temperature have increased at both sites. From 1965-2003, oxygen isotope values on Quelccaya have stayed fairly steady (Figure 4.12), not reflecting either increasing precipitation or temperature at Sicuani and Ccatcca. Overall thermal year trends at Sicuani and Ccatcca are similar to the wet season trends except that Sicuani temperatures have stayed fairly constant over thermal year averages. The lacking data in the 1980’s poses a problem for analyzing
increases in temperature and precipitation, yet others have noted rising temperatures in the tropical Andes. Vuille and Bradley (2000) find a rising temperature trend of about 0.1°C per decade in this region.

![Quelccaya Summit Dome δ¹⁸O 1965-2003](image)

Figure 4.12 Quelccaya Summit Dome annual oxygen isotopic values from 1965-2003. Red line (linear trend over 1965-2003) indicates oxygen isotope concentrations have remained stable over the last 38 years.

Nearby weather stations are useful in that the Quelccaya automated weather station has only been in operation for 4 years, while Sicuani and Ccatcca have an extended period of observance. Over the time period during which Sicuani, Ccatcca and the Quelccaya AWS have all been in operation (2004-2008), summit temperature and accumulation measurements correspond closely with temperature and precipitation at the two sites (Figures 4.13 and 4.14). The agreement between accumulation and precipitation amount is surprising, given that precipitation can vary greatly from site to site in the central Andes due to the convective nature of the moisture systems. Some
periods, such as Nov.-Jan. of 2004/2005 do show a difference in accumulation amount, yet features in the data such as peaks and minimums throughout the wet season remain similar.

Figure 4.13 Monthly accumulation (as measured by height change (m)) on the Quelccaya summit AWS, and monthly precipitation (mm) measured at Station Sicuani and Station Ccatcca.

Figure 4.14 Normalized monthly mean temperature at Quelccaya summit, Station Sicuani and Station Ccatcca.
Temperatures at each site were normalized to account for differing elevations of the weather stations (Figure 4.14). The weather station temperature anomalies seem to be a good fit in describing temperature anomalies on the summit, except for the wet season of 2007 in which they underestimate higher temperatures on the summit. Further years of AWS data would provide useful information to further substantiate Sicuani and Ccatcca as a possible proxy to extend the record of precipitation and temperature on the summit back into the mid 60’s.

A direct comparison of oxygen isotopes from the Quelccaya Summit Dome core and wet season temperatures on an annual basis shows a lack of significant correlation over this time period for either of the two weather stations (Figure 4.15 a,b). Correlation between Ccatcca temperatures and Quelccaya $\delta^{18}$O has an r of 0.05 (p > 0.050), and an r of 0.14 (p > 0.050) with Sicuani temperatures. Although correlations may be slightly stronger between QSD oxygen isotopes and wet season precipitation at Sicuani (r = 0.28, p > 0.050) and Ccatcca (r = 0.21, p > 0.050), they are still not significant (Figure 4.15 c,d).
Figure 4.15 (a) Station Ccatcca (b) Station Sicuani average wet season temperature and Quelccaya annual oxygen isotopes, (c) Station Ccatcca (d) Station Sicuani wet season precipitation and Quelccaya annual oxygen isotopes. Wet season precipitation is considered November-April.
Figure 4.15 continued

(c) Station Ccatca Wet Season Precipitation and Quelccaya $\delta^{18}O$

![Graph showing Station Ccatca Wet Season Precipitation and Quelccaya $\delta^{18}O$.](image)

(d) Station Sicuani Wet Season Precipitation and Quelccaya $\delta^{18}O$

![Graph showing Station Sicuani Wet Season Precipitation and Quelccaya $\delta^{18}O$.](image)
4.2.3 Shallow Cores and Snow Pit Data

Many years of snow pit and shallow core data have been collected from the summit of Quelccaya. These provide additional information and update the record from the time when drilling occurred. This section concentrates on the years 2004-2007 during which the AWS was concurrently running. Snow pits were taken in 2005, 2006 and 2007 and generally go to a depth which covers the past year of accumulation. Shallow cores were taken in 2004, 2006 and 2007 and cover two or more years of accumulation data (Figure 4.17).

![Snow Pit Data Graph]

<table>
<thead>
<tr>
<th>Snow Pit</th>
<th>Average (‰)</th>
<th>Maximum (‰)</th>
<th>Minimum (‰)</th>
<th>Range (‰)</th>
</tr>
</thead>
<tbody>
<tr>
<td>2006</td>
<td>-19.73</td>
<td>-12.07</td>
<td>-25.66</td>
<td>13.59</td>
</tr>
<tr>
<td>2005</td>
<td>-16.72</td>
<td>-11.21</td>
<td>-21.39</td>
<td>10.18</td>
</tr>
</tbody>
</table>

Figure 4.16 δ¹⁸O profile of 2005-2007 snowpits taken on Quelccaya summit. Red line shows average isotopic composition of the annual layer. Snow pit years are plotted in reverse chronological order to complement shallow core data. Table shows average, maximum, minimum and range of δ¹⁸O (‰) for all three snow pit years.
The $\delta^{18}O$ average for 2007 and 2006 (-19.41‰ and -19.73‰ respectively) are very similar, and in comparison 2005 is enriched (-16.72‰) (Figure 4.16). Out of the three years, 2005 average temperatures over the wet season were warmer at Sicuani and Ccatcca (12.6°C and 9.8°C respectively), but at the Quelccaya AWS this wet season was actually recorded as the colder of the three years (-3.87°C). There is a possibility this average was affected by the aforementioned problems with minima temperature recordings in February 2005 through June 2005, so a comparison of further years of snowpits and temperature reading would be useful. Precipitation amounts for the years 2005 and 2006 were similar; both at Sicuani and total accumulation for the AWS, while 2007 showed a slight increase. Ccatcca on the other hand records 2005 as the year with least precipitation and 2006 with the most. With such variability in the data over a period of only 3 years it is difficult to make concrete connections with average isotopic values, precipitation amount and temperature for contemporaneous years.

![Graphs showing δ¹⁸O profile of shallow cores taken in 2004, 2006 and 2007 from the Quelccaya summit.](image)

Figure 4.17 $\delta^{18}O$ profile of shallow cores taken in 2004, 2006 and 2007 from the Quelccaya summit.
Each shallow core contains at least two years worth of accumulation, allowing an analysis of post-depositional changes to average isotopic values over time. During drilling, recorded field notes designate the depths of previous thermal years determined by the dry season dust layer.

<table>
<thead>
<tr>
<th>Shallow Core Thermal Year</th>
<th>year drilled (‰)</th>
<th>buried 1 year (‰)</th>
<th>buried 2 years (‰)</th>
<th>Snow Pit (‰)</th>
</tr>
</thead>
<tbody>
<tr>
<td>2004</td>
<td>-19.45</td>
<td>-17.88</td>
<td></td>
<td></td>
</tr>
<tr>
<td>2006</td>
<td>-20.03</td>
<td>-19.11</td>
<td>2006 -19.73</td>
<td></td>
</tr>
</tbody>
</table>

Figure 4.18 Shallow Core and Snow Pit δ¹⁸O values in (‰). Figure provides an average isotopic value for an annual layer during the year drilled, after that ice layer has been buried for one year, and after that same ice layer has been buried for two year. No shallow core was taken in 2005 while there was not a snow pit taken in 2004.

Figure 4.18 shows isotopic values from successive years of shallow core and snowpit data. For example, in the 2004 shallow core the isotopic value for the ice layer 2004 is -19.45‰ (this would be the top layer of the shallow core drilled in 2004). Because no shallow core was taken in 2005, that layer does not show up again until the 2006 shallow core (when the 2004 layer has been buried for two years). After that layer has been buried for two years it has an isotopic value of -17.88‰.

Years during which both shallow cores and snow pits were collected (2006 and 2007), average isotopic values for the thermal year are comparable (2006: -20.03 and -19.73‰, 2007: -19.44 and -19.41‰). For 2006 there is only a 0.3‰ difference between the snow pit and shallow core measurement, and 2007 a 0.03‰ difference. Using
consecutive pit and shallow core data, the isotopic values in the snow trend toward an enrichment of 0.85‰ (0.62‰ including the 2005 data) per year after burial. The isotopic values of thermal year 2005 are unique in that they first become depleted after a year of burial and then become enriched after two years of burial. This is most likely due to the difficulty in designating the specific point of the dry season layer and where to consider the start of the previous year.

Identifying what portion of this enrichment (if any) is attributable to melting and percolation is key to understanding the isotopic response to these processes. In the future, as melting becomes more commonplace on tropical glaciers at even greater elevations, the isotopic records produced will need to be evaluated with these changes in mind. A study of isotopic change after years of burial from shallow cores taken in the 1970’s on the Quelccaya summit, in comparison to isotopic changes occurring from cores taken after melting has occurred may provide some insight to this problem.

In an attempt to find relationships between isotopic values in the snow and accumulation amount or temperature on the glacier, snow pit isotopic data was compared to AWS accumulation and temperature. Snow pit data is particularly helpful because it provides a high resolution isotopic stratigraphy throughout the thermal year. To make the comparison between the weather station and snow pits, each year was analyzed independently and broken down into monthly segments of accumulation or ablation as measured by the AWS. To do this daily increases (or decreases) were added for each month, creating a net snowfall amount for that month. To create an annual accumulation stratigraphy, months with a net decrease were subtracted from the accumulation of the
previous month. Then, each month which contributed positive accumulation was recalculated as a percentage of total snow accumulation over the thermal year. Once a percentage was obtained, the isotopic stratigraphy was binned accordingly. This process is similar to the compilation of accumulation stratigraphies in Hardy et al. (2003). Results are shown in figure 4.19 (a) and (b).

For the three years of comparison (2005-2007), there seems to be a slight trend in the accumulation data toward depleted isotopes with higher accumulation. But there is large scatter in this data, for example note the range of -18‰ is associated with both low and high precipitation amounts. No strong relationship is evident between temperature and isotopes, yet on an annual scale temperature ranges are small. This relationship may only become apparent when temperature ranges become much larger such as during extreme ENSO events or on the range of glacial to interglacial periods.
Figure 4.19 (a) Scatterplot of monthly binned $\delta^{18}O$ from snowpits taken in 2005-2007 vs. contemporaneous average monthly temperature on Quelccaya summit. (b) Scatterplot of monthly binned $\delta^{18}O$ from snowpits taken in 2005-2007 vs. contemporaneous monthly net accumulation on the Quelccaya summit. Triangles represent data from 2005; circles represent 2006 and squares 2007.
4.3 Regional Scale Records of Precipitation and Temperature

Andean meteorological records are lacking both temporally and spatially in comparison to high latitude stations, owing to their remoteness and extreme climate conditions (Hoffman et al., 2003). Of the records that do exist on the Altiplano, many are incomplete (Baker et al., 2001). This fact reemphasizes the importance of understanding ice core records as a proxy for Andean climate over the last 200 years and beyond, but also the difficulty in doing so.

The Altiplano’s longest continuous time series of precipitation comes from meteorological measurements made at Lake Titicaca, South America’s largest and highest freshwater lake (3812 m asl, 8446 km²) (Figure 4.20).
Since 1914, lake level measurements have been recorded by the Servicio Nacional de Meteorología y Hidrología de Peru, at Puno, Peru (Thompson et al., 1992; Baker et al., 2001). Lake Titicaca is considered a dependable recorder of precipitation amount over large portion of tropical South America because it is a nearly closed basin. Baker et al. (2001) demonstrate that the average rise due to precipitation in the wet season (October to March) is much more variable than the average fall due to evaporation in the dry season, thus substantiating lake level as a good proxy for precipitation amount. Because Lake Titicaca lies just 300 km south of Quelccaya, yet captures precipitation from a large watershed area, it was a natural choice to use in comparison to $\delta^{18}$O levels in both Coropuna and Quelccaya.

Over the span of recorded lake level, oxygen isotope levels in both glaciers corroborate well with lake level rise and fall (Figure 4.21). For the most part, years in which lake level is low $^{18}$O is enriched, and years in which lake level is high $^{18}$O is depleted. On an annual basis both Coropuna and Quelccaya oxygen isotope concentrations show good correlation with Titicaca Lake levels (Coropuna $r = 0.47$, $p < 0.001$; Quelccaya $r = 0.46$, $p < 0.001$). After applying a 3 year running mean to smooth noise in the system, Quelccaya’s correlation greatly increases (3-year running mean $r = 0.70$, $p < 0.001$) while Coropuna correlation stays nearly the same (3-year running mean $r = 0.49$, $p < 0.001$). This correlation with Titicaca lake levels suggests a strong connection with South American precipitation over this time period.
Figure 4.21  Standardized Quelccaya and Coropuna $\delta^{18}O$ annual thermal year values compared to November-April Titicaca lake level flux from 1916-2003.

Tropical Andean ice cores have also shown a very good relationship with Tropical Pacific SSTA. Although the original moisture source for this area is the Atlantic and Amazon Basin, Bradley et al. (2003) found strong a strong relationship between Andean ice cores and Pacific Sea Surface Temperatures. Oxygen isotopes in Sajama, Huascarán and Quelccaya all correlate well with Nino 3.4 SSTs from 1973-1984. Using NOAA’s Oceanic Niño Index (a three month running mean of SST anomalies for the Niño 3.4
region (5°N -5°S, 120°-170°W) based on 1971-2000 average) to extend the comparison for Quelccaya and Coropuna back to 1950 shows this relationship holds well for the last half century (Figure 4.22).

Figure 4.22 Upper graph shows Oceanic Niño Index DJF values (black thin line) and Coropuna annual oxygen isotope values (red thick line) from 1950-2003. Lower graph makes the comparison with Quelccaya annual oxygen isotope values (green thick line) over the same period.

The DJF values of the ONI were chosen to compare to Coropuna and Quelccaya $\delta^{18}O$ values, for this is the height of the wet season on the two glaciers and corresponds to the typical time when SSTA in the equatorial Pacific reach their maximum. Coropuna has a stronger correlation with the SSTA’s ($r = 0.55, p < 0.001$) than Quelccaya ($r = 0.24, p > 0.05$), and initially Quelccaya’s relationship is insignificant. Its $\delta^{18}O$ values follow
the major trends of SSTA’s over a longer time frame, but not as well on a year to year basis. On the other hand, Coropuna’s $\delta^{18}O$ values record annual variation of SSTA’s in the Niño 3.4 region quite well. After applying a three year running mean, Quelccaya’s correlation improves drastically ($r = 0.49, p < 0.001$).

Coropuna may capture annual variation in Pacific SSTA’s better than Quelccaya for a few reasons. One, because the western part of the Altiplano is further away from the moisture source it is more sensitive to ENSO generated precipitation anomalies. While it is possible for Quelccaya to still receive some moisture during a period of westerly wind anomalies (due to its more easterly location which is closer to the moisture of the lowlands), it is less likely for this to happen at Coropuna because of its westerly location. Therefore the more westerly sites tend to show a stronger ENSO relationship (Vuille et al., 2000; Garreaud et al., 2003). Also, Coropuna’s location closer to the Pacific Ocean allows the possibility for Pacific moisture to reach the summit of Coropuna and perhaps have a stronger influence on its isotopes, one that is not diminished by first being transmitted through South American circulation components. Finally, melting on the Quelccaya summit has had the effect of dampening the isotopic extremes and smoothing the overall isotopic signal. This may influence the association between the isotopic signature on Quelccaya and Pacific SSTAs.

Noteworthy in this time series is the thermal year 1973 when neither Coropuna nor Quelccaya capture the SSTA’s in the NINO 3.4 region. 1972/73 was an anomalous El Niño year in that it was warm and wet on the Altiplano rather than warm and dry. Although the isotopes may not be able to portray sea surface temperatures for anomalous El Niño or La Niña years, this type of event is not common. The majority of years are
well recorded by Coropuna and if the overall relationship holds beyond the past 200 years, oxygen isotopes in tropical ice cores may be a valuable proxy for SSTA’s in the Pacific and help to further research on the ENSO cycle.
5.1 Annual and Interdecadal controls on oxygen isotopes

In previous literature, $\delta^{18}O$ of tropical ice cores has been presented as proxy for numerous meteorological parameters and circulation components. Broecker (1997), Baker et al. (2001), Hoffman et al. (2003), among others, consider $\delta^{18}O$ as a proxy for Altiplano precipitation and rainout upstream. Bradley et al. (2003) found a strong relationship between equatorial Pacific SST’s and oxygen isotopes in tropical cores, while Henderson et al. (1999), found a significant relationship between Huascarán isotopic values and zonal wind velocities at 500 hPa. Considering records on centennial and millennial scales, Thompson (2000) and Thompson et al. (2006) find positive relationships between $\delta^{18}O$ of tropical cores and temperature.

It is extremely difficult to detangle each parameter’s influence on the isotopic composition in tropical ice cores, because all parameters vary synchronously on interannual and interdecadal time scales. As stated in previous chapters, the central Andes vary between a warm/dry mode (El Niño) which occurs when equatorial Pacific sea surface temperatures are warmer, and a moist/cool mode (La Niña) which occurs when equatorial Pacific sea surface temperatures are anomalously cold. The upper-level winds also vary accordingly, in that during the moist/cool mode, easterly trade winds are
anomalously strong over the central Andes. Pacific sea surface temperature anomalies (hence ENSO) seem to be the driving force behind the synchronous actions of temperature, precipitation and upper-level wind changes over the Altiplano. A proposed mechanism for the way in which all components work together on an annual time scale is described further in the following paragraphs.

A normal wet season on the Altiplano develops as the easterly tradewinds expand due to a southward shifting Intertropical Convergence Zone (ITCZ), and a decrease in the temperature gradient between the tropics and mid-latitudes. At this time, upper level easterlies (300-100 hPa) prevail over the central Andes and lead to lower level moisture transport up the eastern slopes from the continental lowlands (Figure 5.1). Moisture in the ABL reaches the threshold and deep convection leads to precipitation in the high elevations.

Figure 5.1  Atmospheric conditions of 250 hPa geopotential height and wind, during times when moisture is brought to the Altiplano. The easterlies are a result of a weakened meridional temperature gradient. The overall regression field of the wet season is similar to a typical ENSO cold event (La Niña) (Vuille et al., 2008).
Yet, the strength and direction of the upper level winds is ultimately dictated by sea surface temperatures in the equatorial Pacific. During an El Niño event, the thermocline off the coast of Peru becomes depressed and the waters warm because of weakened oceanic upwelling. This reduces the east-west SST gradient (normally the waters in the east are 4-10ºC colder than in the west) which then reduces the pressure gradient. A weakened pressure gradient across the Pacific leads to weakened tradewinds (Cane, 2005). Weakened easterlies create weakened upslope flow on the eastern slopes of the central Andes and less moisture transport from the continental lowlands, creating a drier than normal wet season (Garreaud et al., 2003). The opposite is true for a wet season which coincides with La Niña events. The easterlies are strengthened due to cooler waters off the coast of Peru and moisture is brought up the eastern slopes more consistently.

Although these parameters vary together on an interannual and interdecadal time-scales, events in which portions of the cycle become decoupled are evident over the last 100 years of meteorological observation. Examples include the 1972/73 El Niño year in which the Altiplano experienced warm but wet conditions. And the 1988/89 La Niña year brought cold but dry conditions to the central Andes (Garreaud and Aceituno, 2001). Upper level winds are very sensitive to the exact position and intensity of the sea surface temperature anomalies in the central equatorial Pacific. Not all ENSO events are the same, and each one will vary as to the way the SST anomalies develop in the equatorial Pacific, which in turn changes the placement of anomalous upper level winds in the central Andes.
5.2 Assessing the Longer Time-scale

There are also cases on the longer term that may not fit into the mold of a warm and dry or cold and wet Altiplano as well. If warmer temperatures and drier conditions (or colder temperatures and wetter conditions) have not always varied synchronously in the past, what controls the isotopic signature over these longer time scales?

Thompson et al. (2006) created an Andean composite which gives a view of decadal δ¹⁸O averages over the last 2000 years (Figure 5.2). Two important aspects of this record are enriched isotopes from AD 1100-1300 (during the Medieval Warm Period), and depleted isotopes from AD 1400-1900 (during the Little Ice Age) The greatest depletion occurs from AD 1810 to 1812, signaling strong cooling that was produced by large volcanic eruptions in 1809 (unknown) and in 1815 (Tambora) (Thompson et al., 2006).

Figure 5.2 Composite of three Andean δ¹⁸O ice core records, decadally averaged over 0-2000 AD (Thompson et al., 2006).
An important advantage of using ice core records as proxy data for tropical climate is that they have numerous parameters which record climate events and some core histories (such as Huascarán and Sajama) extend back to the Last Glacial Maximum. Along with $\delta^{18}O$, dust concentrations, aerosol chemistry, and accumulation rates can give a fuller picture of past climate conditions (Thompson et al., 2006).

Because the Quelccaya record shows well-defined annual layers for the past 400 years, it was possible to create a net mass accumulation ($A_n$) history over this period (Figure 5.3) (Thompson et al., 2006). Both Sajama and Huascarán’s rapidly thinning ice layers (with depth) prohibited the construction of an accumulation history for those sites, so Quelccaya remains the sole reliable $A_n$ history for the Andean region. Yet a high resolution pollen history from Sajama over the same time period provides support for the conditions described by reconstructed accumulation history of Quelccaya (Thompson et al., 2006).

![Figure 5.3 5-yr averages of $\delta^{18}O$ (‰) (left), and 5-yr averages reconstructed Quelccaya accumulation ($A_n$) (right) back to A.D. 1600 (Thompson et al., 2006)](image)

This reconstruction history shows a clear distinction between a wet first half of the Little Ice Age and a drier second half. Interestingly, isotopes over this time period
remain consistently depleted (Thompson et al., 2006). If precipitation was the dominant controlling factor over this time period, then the oxygen isotopes should reflect this pattern.

Looking farther back into the record at relationships over centuries and millennia provides additional evidence of the positive correlation between temperature and isotopes in the long term. Oxygen isotopes in polar ice cores have long been recognized as a proxy for surface temperature. Late Glacial Stage ice from Dome C, Antarctica shows 6‰ depletion (across the glacial-interglacial transition) and the GISP2 core from Greenland shows 5.1‰ depletion over this period. Interestingly, a depletion of 6.3‰ is recorded in LGS ice relative to Holocene ice in the tropical Huascarán core (Figure 5.4), an amount of depletion very similar to the polar cores (Thompson, 2000).

![Figure 5.4 100-yr averages of δ¹⁸O, nitrate, and dust from Huascarán Core 2 over the last 20,000 years. Large depletion of oxygen isotopes over glacial-interglacial period is apparent (Thompson, 2000).](image)
Also evident in the isotopic profile is a discrete climate reversal at the same time as the Younger Dryas cold phase prominent in the North Atlantic region (Thompson, 2000). The Sajama δ¹⁸O profile (Figure 5.5) also shows a strong depletion of 5.4‰ over LGM ice and indicates a climate reversal around the time of the Younger Dryas.

![Figure 5.5 100-yr averages of δ¹⁸O, nitrate and dust from Sajama Core 1 over the last 25,000 years. Depletion of oxygen isotopes over glacial-interglacial period is apparent (Thompson, 2000).](image)

Although their overall isotopic depletion on glacial-interglacial timescales is similar, these two sites record different moisture settings. Huascarán contains higher concentrations of dust during LGS, evidence that the environment in the tropics (9ºS) was drier at this time. On the other hand, large regional paleolakes existed near the location of Sajama in the subtropics (18ºS) and its record accordingly does not contain elevated dust levels (Thompson, 2000). It is difficult to reconcile the similar isotopic depletion and different moisture environments while considering a precipitation driven oxygen...
isotope signal on millennial time scales. Before oxygen isotopes in tropical ice cores can be used as a paleo-precipitation proxy in the central Andes it is imperative this issue (along with evidence from the LIA) is addressed.

5.3 Trends Over the Last Century

In comparing temperature and precipitation on Sajama to weather station data and NCEP temperature data, Hardy et al. (2003) found a statistically significant (negative) relationship between $\delta^{18}O$ and precipitation amount, and a weak, not significant positive relationship with NCEP temperature. Yet the results in Chapter 4 only show a weak relationship between oxygen isotopes and accumulation on a seasonal basis, while other cases (annual relationship between $\delta^{18}O$ and precipitation at Sicuani and Ccatcca) show no significant correlation at all. The snowpit $\delta^{18}O$ profiles as compared to accumulation seem to show a weak (albeit quite scattered) trend toward isotopic depletion with increased accumulation on the summit. Yet it would be hard to develop any type of isotopic depletion rate per unit accumulation because the variance is so great. For example, similar isotopic values of -17.15‰ and -17.10‰ from the monthly binned snowpit data, correspond to monthly accumulation amounts of 0.09 m (March 05) and 0.47 m (December 2004) accordingly. These accumulation amounts are examples of the low end and the high end of the spectrum, yet have very similar isotopic values. Conversely, similar accumulation rates also show quite a difference in isotopic depletion. This is not to say that with further years of snowpit data, a more tenable relationship could not be found.
The temperature relationship is difficult to interpret because over the period of the wet season temperature changes are quite small. Even within the 2°C temperature range similar isotopic values span the end member values of temperature. Some confounding factors complicate this matter. Measuring temperature at the summit is not the same as measuring temperature at the level in the atmosphere where condensation of water vapor actually occurs. However, horizontal temperature variations in the tropics are relatively small above the thin atmospheric layer above the Earth’s surface (Thompson et al., 2006), so this may be of minimum consequence. Also, temperature as measured on the summit, should be near to those of free atmospheric values, but this is not always so.

Results from comparisons between Quelccaya isotopic values and temperature from the weather stations Ccatcca and Sicuani were also scattered and no significant trends were apparent. But because data from the two weather stations seem to agree well with precipitation amounts and temperature anomalies on the summit of Quelccaya, a longer record of conditions on the summit may be possible. This would be interesting in its own right since high elevation temperature and precipitation records are nearly nonexistent, or only span a period of a few years. Further years of data from the Quelccaya AWS would be essential to this endeavor.

Although temperature range over the wet season is minimal, temperature changes on longer time scales (e.g. centennial, glacial/interglacial) can be more apparent and could play a larger role. This seems to be the case for time periods such as the LIA and LGM.
The strongest relationships discovered by the explorations of numerous regional and local conditions in this thesis occur on more regional scales over longer time periods. Both equatorial Pacific sea surface temperature anomalies and Lake Titicaca levels show a strong relationship with isotopes.

Yet as Thompson et al. (2006) and Thompson (2000) have shown, strong positive temperature relationships do not appear on time-scales much less than a century long. The relationships in this thesis have been based on comparing annual differences and similarities between these different climate records, yet actual trends over the last 100 years may be different. A look at the overall trends in Lake Titicaca levels during the period of measurement, and isotopes in both Quelccaya and Coropuna show a different relationship than has been seen by comparing annual differences (Figure 5.6).

(a)

![Titcaca Lake Level 1916-2000](image)

Figure 5.6 Trends in (a) Titicaca Lake levels, (b) Quelccaya and (c) Coropuna oxygen isotope values over the period 1916-2000 AD. Red line indicates trend over 1916-2000 AD.
As Titicaca Lake levels have generally been increasing over 1916-2000, both Coropuna and Quelccaya have shown enrichment of $\delta^{18}O$. If Lake Titicaca levels are in fact increasing due to increased precipitation, then a factor other than rainout is controlling the isotopic enrichment.
Determining the likelihood of increased precipitation on the Altiplano over the last 100 years is difficult. Only two other records on the Altiplano, La Paz, Bolivia (on the southern end of the basin), and the Quelccaya net accumulation record, span this similar time period.

Precipitation has been measured at La Paz since 1898 (WMO station code 852010). The time period from 1916 until 2004 at La Paz shows a quite different story than the Titicaca Lake levels (Figure 5.7). Over the past century, the trend in precipitation is decreasing at La Paz. This is interesting because precipitation at La Paz and Lake Titicaca water levels have been well correlated in the past. Hastenrath (2004) found a strong positive correlation ($r=0.66$, significant at 1% level) from 1915-1968 between lake levels and precipitation at La Paz.

![Figure 5.7 Trends in total annual precipitation at La Paz, Bolivia. Red line indicates trend over 1916-2000 AD.](image)
The Quelccaya reconstructed net accumulation data seems to agree with declining precipitation on the Altiplano over the last century (Figure 5.8). Although the strength of the signal is not as strong in the accumulation data as in the La Paz precipitation record, accumulation does show a slight decreasing trend.

![Quelccaya Reconstructed Net Accumulation 1916-2000](image)

**Figure 5.8** Reconstructed net accumulation (in m w.e./year) from the QSD 2003 core. Red linear trend line indicates a slight decrease in accumulation on Quelccaya over 1916-2000.

Using just two records does not prove that precipitation trends have been decreasing on the Altiplano, yet it is a good indication. If precipitation trends have been decreasing over the last 100 years (while temperatures have been increasing), then these two parameters can still be considered in synch over the last century (the warm/dry Altiplano mode). Although this does not provide insight into periods that may have been cold and dry or warm and wet over longer timescales, it is evidence of the dominance of the 2 mode climate system over the past 100 years on the Altiplano.
Lake Titicaca has long been considered an excellent proxy for precipitation on the Altiplano. Because it occupies a large watershed, the sediments portray a regional incorporated moisture budget (Cross et al., 2000). Baker et al. (2001) have interpreted lake level rise to indicate wet conditions and lake level fall to indicate dry conditions. Melice and Roucou (1998) used Lake Titicaca as a basin wide rain gage for their interpretations, while Fritz et al. (2007) found that periods of glacial advance in the southern tropical Andes occurred during the same time as high lake levels and periods of glacial retreat occurred during times of negative water balance.

Conditions now are showing glacial retreat and rising lake levels. It is quite possible that glacial melt is influencing Lake Titicaca water levels. The basin itself is surrounded by mountains of the eastern and western cordillera, which currently hold many glaciated peaks. Yet all of the glaciers that have been researched are currently retreating (Fritz et al., 2007; Francou et al., 2003). Lake levels are extremely sensitive to the precipitation/evaporation balance. Present-day hydrologic inputs to Lake Titicaca consist of 47% direct rainfall and 53% inflow from 6 major rivers. Evaporation accounts for 91-99% of water loss, while the remaining loss happens through discharge from the only surface outlet of the Lake, Rio Desaguadero (Abbott, et al., 1997; Reference within Fritz et al., 2007: Roche et al., 1992). Large changes in the precipitation/evaporation balance would not be able to explain rising lake levels if precipitation is indeed decreasing. The most logical explanation is increasing inflow from rivers which would include glacial melt.

These results hold implications for using Lake Titicaca and the connecting basins as indicators of precipitation trends on longer timescales. Baker et al. (2001) concludes
that “Lake Titicaca is a reliable recorder of the precipitation that falls on a large portion of tropical South America.” And while this may be true on an annual basis, looking at anything longer (such as centennial or millennial scales of rise and fall) may give an incorrect impression during warm times when glacial melt is large. Repercussions of this would only be a problem further back in time when lake cores can only be centennially or millennially resolved.

Both Quelccaya and Coropuna show a definite enrichment of oxygen isotopes over the last century. This is in agreement with both increasing temperature and decreasing precipitation over this time. The isotopic profile of these two places communicates a strong regional and large scale integrated climate scenario. Relationships with singular or similarly located stations are insignificant (Sicuani and Ccatcca), while records that incorporate regional scale changes are strong (e.g. Titicaca and equatorial Pacific SSTA’s).

5.4 Future Directions

Future research should address periods of extreme temperature change and isotopic response. It may be useful to concentrate on the largest El Niño and La Niña events and the temperature response at all levels in the tropical troposphere. Since it has been hypothesized that extended periods of El Niño or La Niña style climate may have existed in the past (Bradley et al., 2003), concentrating on extreme ENSO events may be analogous to extended warm or cold periods on the Altiplano. Also, investigations into periods of past climate on the Altiplano during which 4 possible modes of climate existed (warm/dry, warm/wet, cold/dry, cold/wet) would be useful. In further studies,
importance would lie in taking a multi-proxy approach as well as using all variables of ice core data in concert. For example, analysis of dust records and isotopic profiles together provide a clearer view of the climate system. Previous research has concentrated on isolating oxygen isotopic response to climate dynamics, yet this is a very limited view.

Although these relationships are still under debate, this does not negate the important message from the isotopic profile. It is quite clear that large changes in climate dynamics have occurred in the past, and current changes are underway.
LIST OF REFERENCES


Thompson, L.G., Mosley-Thompson, E., Dansgaard, W., Grootes, P.M., 1986. The "Little Ice Age" as recorded in the stratigraphy of the tropical Quelccaya ice cap, *Science*, 234, 361-364.


