AN ABSTRACT OF THE THESIS OF

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Abstract approved:

Anne W. Nolin

Declines in glacier area and volume are widespread. These changes will have important hydrologic consequences since glaciers store tremendous amounts of fresh water and buffer seasonally low flows in many densely populated regions. In this thesis I focus on a region that is hydrologically vulnerable to glacier change, namely the Cordillera Blanca, Peru. I present three manuscripts that focus on measuring glacier area change, modeling the effect of this area change on the hydrology of one watershed, and isotopic sampling to elucidate hydrologic processes in this watershed and the entire Cordillera Blanca.

In the first manuscript, I describe a methodology for mapping glaciers using satellite imagery. Satellite data, in conjunction with automated glacier mapping methods, are being used more frequently to map changes in glacier size. In contrast to the majority of studies using automated methods, I correct satellite images for atmospheric effects. Mapping glaciers with atmospherically-corrected satellite images resulted in an approximately 5% increase in glacier area, relative to glaciers mapped with non-atmospherically-corrected images. I also applied a consistent threshold that was validated using high-resolution satellite imagery. This helps to reduce error associated with change analysis. For the entire Cordillera Blanca, I calculated a 25% decrease in glacier area
from 1987 to 2010. The rate of glacier area loss has increased significantly based on the most recent estimates.

In the second manuscript, I use a physically-based, hydrologic model, the Distributed Hydrology Soil Vegetation Model (DHSVM) with a newly-coupled dynamic glacier model to simulate stream discharge and glacier change in the Llanganuco watershed of the Cordillera Blanca. I also examined statistical trends associated with historical records of temperature, precipitation, and discharge. I observed significant positive trends in annual temperature, but no trends in precipitation or discharge despite a 25% reduction in glacier area in this watershed over the same time. The model setup process and the results of sensitivity analyses are described. Of the input parameters I examined, I found that the model was particularly sensitive to changes in albedo and precipitation. Based on established efficiency criteria, the newly-coupled model did a decent job of simulating historical stream discharge and glacier area during 10 year calibration and validation periods. However, due to the lack of additional validation data and an inability to quantify uncertainty associated with model output, the model is not yet ready to be used for predicting future discharge based on different climate projections.

In the third manuscript I describe the knowledge gained about hydrologic processes from isotopic sampling in the Llanganuco watershed, as well as other watersheds of the Cordillera Blanca. Thirty water samples from Llanganuco were collected in July 2011 and measured for stable isotopes of water, δ\textsuperscript{18}O and δ\textsuperscript{2}H. I first calculated the isotopic lapse rate, or the relationship between isotopic values and elevation. Lapse rates from this watershed are slightly more positive than global averages. This observation is best explained by the influence of glaciers. I also calculated the strength of the relationship between isotopic values and percent glacier cover. For Llanganuco, glacier cover is a better predictor of isotopic value than elevation. Based on examination of the same relationships at larger scales in the Cordillera Blanca, this relationship appears to be persistent at a regional scale. Finally, I used a simple two-component mixing model to estimate the relative contributions of glacier meltwater and groundwater in the
Llanganuco watershed. Glacier meltwater made up approximately three-fourths of surface water that exited the watershed during this two week period in July, 2011. The importance of glacier meltwater is clearly demonstrated using stable isotopes, but further, more detailed monthly sampling is necessary to accurately determine annual and dry season streamflow contributions from glacier meltwater and groundwater.
Glacier Change in a Basin of the Peruvian Andes and Implications for Water Resources

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Patrick J. Burns

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I understand that my thesis will become part of the permanent collection of Oregon State University libraries. My signature below authorizes release of my thesis to any reader upon request.

_____________________________________________
Patrick J. Burns, Author
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I would like to start off by thanking my advisor, Anne Nolin. She has challenged me since I first arrived at Oregon State University. Anne encourages me to think big, but also to pay attention the details and understand physical processes. I thank Anne for giving me every opportunity to succeed and for selecting me to work on such an interesting and meaningful project. I’d like to thank the rest of the Mountain Hydroclimatology Research Group for all of their feedback and suggestions during group meetings and paper reviews. Kelly Gleason deserves special thanks for serving as my field assistant in Peru. Kelly’s knowledge of the Spanish language and tough field demeanor were very helpful.

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I have been working on the modeling portion of this project with Chris Frans and Bibi Naz from the University of Washington. Chris and Bibi have answered countless questions about the hydrologic model we used in this study. Their expertise in programming has been very valuable, helping me to realize the importance and necessity of this skill.

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I need to thank Regine Hock for organizing one of the best experiences of my graduate education, the International Summer School in Glaciology which was held in McCarthy, AK. Following this field class, I was lucky enough to attend the annual International Glaciological Society (IGS) Symposium. Both the course and the conference were tremendous experiences which helped to advance my thesis.

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CONTRIBUTION OF AUTHORS

Manuscript 1: Dr. Anne Nolin supervised this project and helped to develop the methodology.

Manuscript 2: Dr. Bibi Naz and Chris Frans performed most of the programming for the model and also gave me technical guidance. Dr. Anne Nolin provided feedback during the modeling process. Dr. Garry Clarke developed the glacier model and assisted with its implementation. Dr. Dennis Lettenmaier supervised the project. Dr. Bryan Mark provided data for the modeling study and helped to collect data in the field. Dr. Thomas Condom provided hydrologic data for the study area.

Manuscript 3: Dr. Renee Brooks and Dr. Anne Nolin helped to develop the methodology. Dr. Brooks provided substantial feedback during the writing process. Dr. Bryan Mark shared isotopic data from samples collected from 2004 to 2011. Ryan Gordon and Kelly Gleason helped to collect samples and interpret results.
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1 Introduction

1.1 Motivation and goals of this thesis

Nearly 70% of the Earth’s fresh water is stored in permanent snow, ice caps, or glaciers. Glaciers are sometimes referred to as water towers (Viviroli et al., 2007) because they store vast quantities of water in its frozen form. In many regions, this storage compensates streamflow during drier periods. Over approximately the past half-century glacier change has become increasingly more relevant for three primary reasons. First, glacier recession has alarming implications for sea level rise (Berthier et al., 2010; Jacob et al., 2012; Nicholls and Cazenave, 2010). This is because the volume of water equivalent lost by a glacier is added to streamflow, the groundwater reservoir, or the atmosphere, but ultimately ends up in the oceans causing global sea level to rise. Second, glaciers are sensitive indicators of how the Earth’s climate is changing, especially in the tropics (Kaser and Osmaston, 2002; Thompson et al., 2011). Glaciers are particularly sensitive to changes in temperature and precipitation, and relatively small shifts in either of these quantities can have large effects on the mass balance of a glacier. Third, glacier change has become increasingly relevant because glacier melt is a vital component of local and regional water resources, especially in regions where there are seasonal distinctions between temperature and/or precipitation. In these places, glacier meltwater buffers seasonally low flows and provides a greater overall specific discharge relative to nonglacierized basins (Fountain and Tangborn, 1985; Mayo, 1984). In a number of regions, glaciers are situated near zones of high population density (Figure 1.1) and the implications of their recession are potentially dire (Barnett et al., 2005).

Air temperature increases have been widespread across the globe and observed decreases in snow and ice are consistent with this warming (Lemke et al., 2007) (Figure 1.2). However, air temperature increases cannot explain observed glacier change everywhere. Depending on the location, changes in other components of the surface energy balance, as well as precipitation, sometimes explain the observed glacier changes better than
temperature alone. Glacier retreat and mass loss have been well documented around the world, most notably in mountain ranges adjacent to highly populated regions, including Northwestern North America (Bolch et al., 2010; Moore et al., 2009), the Hindu Kush-Himalaya (Bolch et al., 2012; Committee on Himalayan Glaciers et al., 2012; Kaab et al., 2012), and the Tropical Andes (Alvaro et al., 2009; Racoviteanu et al., 2008; Vuille et al., 2008). While field-based glacier mass balance measurements are traditionally considered to be the most representative of glacier health, observations from remote-sensing instruments are often more practical because they have higher temporal and spatial coverage for mountainous areas which are usually difficult to access on the ground.

Automated methods for mapping glaciers with satellite imagery can significantly improve analyst processing times, relative to hand digitization, and are more precise (Paul et al., in press). These methods rely on the difference in snow and ice reflectivity measured in the visible or near-infrared and the mid-infrared portions of the electromagnetic spectrum. A threshold value is applied to segment the non-glacier and glacier ice areas (Pellikka and Rees, 2010). However, some studies of glacier change fail to use a consistent, validated threshold value. Furthermore, the vast majority fail to account for atmospheric effects, and instead use raw data measured at the satellite. These issues are problematic because they can introduce additional error into area change estimates.

It is important to quantify the extent of glacier change and the uncertainty associated with these estimates so we may relate these changes to observed discharge and then make predictions about future flows. Many have predicted that water resources in mountainous regions will be negatively impacted by continued climate warming (Barnett et al., 2005; Immerzeel et al., 2010; Juen et al., 2007). In the tropical Andes, these effects seem to be inevitable as air temperature is projected to increase more at higher elevations (Bradley et al., 2006). When it comes to snow and glaciers in the tropical Andes, the most pressing problem is a decrease in water storage. In the Cordillera Blanca, Peru, 90% of precipitation occurs from October to April. Dry season streamflow is substantially buffered by glacier melt (Mark and McKenzie, 2007). If glaciers disappear then the water
they once stored will no longer be available. Ultimately then with continuing glacier retreat, we would expect the runoff regime in the Cordillera Blanca to become more and more similar to the precipitation regime which has a very strong seasonality (Juen et al., 2007).

In this study we focus on the Llanganuco watershed (centered at -9.05° S, -77.61° W) of the Cordillera Blanca, Peru. The watershed has an area of 89 km², approximately one-third of which is glacierized, and drains into the northwest-flowing Rio Santa, which is part of the larger regional watershed called the Callejón de Huaylas. Based on 2007 Peruvian census data, this watershed has an estimated 267,000 inhabitants (Mark et al., 2010). In this semi-arid region, water originating from snow and ice is vital for drinking water, agriculture, and hydropower, which generates an estimated 80% of Peru’s electricity (Vergara et al., 2007). With a large population that partially depends on receding glaciers, water vulnerability is high in this region (Bury et al., 2011) prompting a need for accurate predictions of glacier contributions to streamflow.

The stable isotopes of water (δ¹⁸O and δ²H) are useful tools for distinguishing between different water sources (Clark and Fritz, 1997). Previous studies performed in the Cordillera Blanca have used stable isotopes to determine relative contributions of glacier meltwater and groundwater in a glacierized watershed (Baraer et al., 2009), as well as to show an increase in specific discharge from glacierized watersheds (Mark and McKenzie, 2007). In order to fully understand the hydrologic implications associated with glacier change, the current contribution of glaciers to streamflow needs to be accurately quantified. Much more work needs to be done before stable isotopes can be used do accurately estimate the seasonal and annual contributions of glacier meltwater to streamflow in the entire Callejón de Huaylas. In order to make these estimates, isotopic variation of the different source waters in the Cordillera Blanca will need to be explained as a function of time and space.

In response to the issues outlined above, we ask the following research questions:
1. How has glacier extent in the Cordillera Blanca changed from the beginning of the satellite record (~1975) to present?

2. How important is the application of an atmospheric correction procedure to the accuracy of a glacier area estimate derived from satellite imagery?

3. When using automated glacier mapping methods, how important is the choice of threshold for estimating glacier area?

4. How well does a new distributed, glacio-hydrological model perform in a test watershed of the tropical Andes?

5. How do the stable isotopes of water vary spatially during the dry season, and what is the primary explanation for this variation?

6. What is the fractional contribution of glacier meltwater to dry season streamflow in a watershed of the Cordillera Blanca?

With a focus on the Cordillera Blanca and the Llanganuco watershed we present the following three manuscripts to answer the research questions outlined above. The first paper, Using atmospherically-corrected Landsat images to measure glacier area change in the Cordillera Blanca, Peru, focuses on quantifying the extent of glacier change in the Cordillera Blanca from 1987 to 2010. Unlike most other studies which utilize satellite data to estimate glacier area loss, we used four atmospherically-corrected satellite scenes to make our estimates. We quantified these area changes between dates and as a function of space. A consistent, validated glacier mapping threshold was used to minimize error.

The second paper, Distributed modeling of runoff in a glacierized basin of the Peruvian Andes, details the use of a newly coupled, distributed glacio-hydrological model in the Llanganuco watershed. We describe the setup of the model and selection of model parameters. We used bias-corrected reanalysis data to run the model. The model was calibrated using measured streamflow as well as glacier area estimates from the first manuscript. Trend analyses were also performed on historical records of temperature, precipitation, and discharge to determine the effect that the observed glacier area loss has had on streamflow.
The third paper, *Isotopic variation during the dry season in a glacierized watershed of the Peruvian Andes*, focuses on the spatial variability of the stable isotopes of water in Llanganuco. We collected 30 water samples in July 2011 from different source waters and examined how they varied with median subwatershed elevation and percent glacier cover. These relationships were then examined at larger scales in the Cordillera Blanca to determine if they were persistent. Finally, a two-component mixing model was applied to estimate the relative contribution of glacier meltwater to streamflow. This estimate serves as a useful check on the model output from the second manuscript.

1.2 Figures

**Figure 1.1:** Population densities estimated for 2010 in people per km\(^2\) (Center for International Earth Science Information Network (CIESIN) et al., 2005) and locations of glaciers taken from the Randolph Glacier Inventory (Arendt et al., 2012).
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Using atmospherically-corrected Landsat images to measure glacier area change in the Cordillera Blanca, Peru

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2.1 Abstract

The dynamic, tropical glaciers of the Peruvian Cordillera Blanca are rapidly changing and these changes will affect water availability, especially during the dry season. In this study, we quantify recent changes to these water reservoirs, providing estimates of glacier area in the Cordillera Blanca and sub-watersheds of the Rio Santa for the following years: 1987, 1996, 2004, and 2010. To map glacier area change we used high-resolution satellite imagery to calibrate and validate our selection of a single threshold for the Normalized Difference Snow Index (NDSI). This threshold value was applied to all NDSI images, which were derived from four atmospherically-corrected Landsat TM scenes acquired at the end of the local dry season. We determined that debris-free glacier area estimates are sensitive to the choice of threshold. We also explored the effects of atmospheric correction by comparing debris-free glacier area estimates generated using atmospherically-corrected and -uncorrected scenes. Our results suggest that atmospheric correction can have a significant impact on debris-free glacier area estimates. Debris-free glacier area estimates derived from uncorrected scenes are approximately 5% less than debris-free glacier area estimates derived from atmospherically-corrected scenes. In order to calculate total glacier area we manually mapped debris-covered glaciers, because automated methods were unsuccessful in this region. As of August 2010, the Cordillera Blanca had a total glacier area of 482 km², which amounts to a 25% decrease since 1987. Glaciers in the southern portions of the Cordillera Blanca, which have lower median elevations on average, lost a greater percentage of their area from 1987 to 2010, relative to their northern counterparts. Overall, glacier area loss in the Cordillera Blanca appears
to be accelerating: between 2004 and 2010 glaciers in the Cordillera Blanca lost area at a rate that was approximately 3.5 times the average rate of area loss from 1970 to 2003.
2.2 Introduction

Tropical glaciers, like those in the Cordillera Blanca of Peru, are sensitive indicators of climate change (Kaser and Osmaston, 2002) and vital dry season sources for drinking water, agriculture, and hydropower generation (Bradley et al., 2006; Mark and McKenzie, 2007). Recent studies focusing on hydrologic modeling of glacier contribution to watersheds in the Cordillera Blanca have utilized multi-temporal estimates of glacier area derived from remotely sensed data to describe the effect that glacier change has had and will have on water resources for this region (Baraer et al., 2012; Condom et al., 2011; Juen et al., 2007). Remote sensing studies focusing on this region have shown that multispectral satellites such as Système Pour l'Observation de la Terre (SPOT) (Georges, 2004; Racoviteanu et al., 2008), Advanced Spaceborne Thermal Emission and Reflection Radiometer (ASTER) (UGRH, 2010), and Landsat Thematic Mapper (TM) (Mark and Seltzer, 2003; Silverio and Jaquet, 2005) are useful tools for monitoring changes in glacier extent on approximately decadal time scales. It is crucial that the shrinking water reservoirs of the Cordillera Blanca continue to be accurately monitored both on the ground and remotely since meltwater makes up a significant portion of dry season discharge (Mark et al., 2005).

Manual delineation, or simple hand-digitization, of remotely sensed images had been considered to be the most accurate method for mapping glaciers (Albert, 2002), but this method is very time-consuming for a multi-temporal change analysis of a large area, such as an entire mountain range. Automated glacier mapping methods using various band ratios have been used to map glaciers at local to regional scales because these methods are easier to implement and frequently more accurate than manual digitization (Paul et al., in press). Automated methods for mapping glacier ice utilize different bands of the electromagnetic spectrum. Debris-free glacier ice is most reflective in the visible part of the electromagnetic spectrum (0.4-0.7 μm), less so in the near infrared (0.7-1.0 μm) portions of the electromagnetic spectrum, and far less reflective at longer wavelengths. It is this contrast in reflectance that enables automated mapping of debris-free glacier ice. Automated methods that rely on this contrast in reflectance are not effective for mapping...
debris-covered glaciers because the rock debris obscures the underlying ice. While it is important that both debris-free and debris-covered ice are mapped accurately to account for total glacier area, the major focus of this paper is mapping changes in debris-free glacier area.

Paul et al. (2007) found that a simple band ratio using Landsat TM bands 3 (0.63 – 0.69 µm) and band 5 (1.55 – 1.75 µm) was effective for mapping shadowed ice, but tended to misclassify water bodies as ice. A simple ratio using Landsat TM band 4 (0.76 – 0.90 µm) and band 5 has also been used effectively for mapping glacier ice (Jacobs et al., 1997; Paul et al., 2002), but this method is less effective in deeply-shadowed areas. Racoviteanu et al. (2008) used the Normalized Difference Snow Index (NDSI) (Hall et al., 1995) with data from SPOT to map glaciers in most of the Cordillera Blanca, showing that the NDSI was effective at distinguishing glacier ice from non-ice areas, especially where shadowed ice is a problem. The NDSI uses the difference between a visible and mid-infrared band divided by the sum of those bands. The NDSI is normally highest over fresh snow and lowest over wet and/or dirty ice.

Even though most mountain glaciers are at high elevations and, relative to targets near sea level, there is less atmospheric mass between the target and the sensor, optical satellite imagery used in multi-temporal glacier change studies should be atmospherically-corrected to account for atmospheric scattering by gaseous and aerosol constituents. Atmospheric scattering and absorption affect light transmittance through the atmosphere and distort the measured reflectance characteristics from surface materials. Reflectance from the target of interest is modified by atmospheric effects, and path radiance, which is light that has not interacted with the target, will also reach the sensor (Vermote et al., 1997). The process of absorption, either by gas molecules or terrain, converts the sun’s energy to a different form. Most satellite sensors have been optimized to record electromagnetic data from atmospheric windows where atmospheric absorption is less. For mountain glaciers, the process of Rayleigh scattering is particularly important because most Rayleigh scattering, commonly by gas molecules like oxygen and nitrogen,
takes place between 2 and 8 km in the atmosphere (Jensen, 2005). Atmospheric correction is especially important for mapping methods which utilize bands in the visible spectrum, such as the ratio of TM band 3 divided by TM band 5 or the NDSI, because atmospheric Rayleigh scattering varies inversely to the fourth power with wavelength (Cracknell and Hayes, 1991). Mie scattering, or aerosol scattering, usually occurs in the lower 4.5 km of the atmosphere while non-selective scattering of water vapor usually takes place in the lower 2 km of the atmosphere. Nearly all remote sensing studies focusing on glacier change do not perform atmospheric correction and instead use raw top-of-the-atmosphere data. The few studies that performed atmospheric correction, usually implemented a relative atmospheric correction by subtracting the darkest pixel value in a scene, commonly referred to as Dark Object Subtraction (DOS) (Albert, 2002).

Another concern associated with automated glacier mapping methods is that they often use varying thresholds for distinguishing between ice and non-ice surfaces. There has been no consensus from glacier remote sensing studies on how to choose the appropriate threshold for mapping glacier ice with multispectral satellite data and whether or not a threshold should be consistent from scene to scene. Ideally in a change analysis, in order to minimize error, a threshold should be consistent if an analyst is using data from the same sensor at approximately the same time of year for multiple years. Atmospheric correction should help to facilitate the selection of a single threshold since it essentially standardizes scenes from different dates, with the assumption that other sources of variation, such as smoke, thin clouds, or fresh snow, are not introduced.

A final concern is that the choice of a threshold for automated glacier mapping methods is often not validated. Objective testing and application of validated thresholds is important for consistent inter-annual and cross-regional comparisons of glacier areas. Many studies offer little explanation as to how a threshold value was selected (if the threshold value is even stated). Some studies make visual comparisons between glacier extent mapped with a certain threshold and a color- or false-color-composite image from the same scene (Paul and Andreassen, 2009). Others choose a threshold based on an
inspection of the NDSI image histogram (Silverio and Jaquet, 2005). When possible, a threshold choice should be validated using either ground data or higher-resolution satellite imagery from the same date.

In this study we use atmospherically-corrected Landsat TM imagery and a single, validated glacier threshold to provide an accurate, consistent measure of glacier change in the Cordillera Blanca, Peru. We examine glacier area change for 11 sub-watersheds of the Rio Santa (Figure 2.1), focusing on those with long term discharge records. We also estimate the total glacier area change for the entire Cordillera Blanca to fill in data gaps and compare our estimates with previous studies. To illustrate the effect of atmospheric correction we focus on one particular watershed, Llanganuco, because of data availability at this location. The specific objectives of this study are to: (1) choose and validate a single threshold for mapping glaciers in the Cordillera Blanca; (2) compare estimates of Llanganuco debris-free glacier area generated from atmospherically-corrected satellite imagery and atmospherically-uncorrected imagery; and (3) to quantify glacier area change from 1987 to 2010 in gauged watersheds draining to the Rio Santa, as well as the entire Cordillera Blanca.

2.3 Study Area

The Cordillera Blanca contains the highest concentration of glaciers anywhere in the tropics (Kaser et al., 1990). Previous work has shown that glacier area in the Cordillera Blanca decreased by 15-22% from 1970 to 2003 (Georges, 2004; Racoviteanu et al., 2008). The range depends on the value used for the 1970 glacier area. More recently, a thorough glacier inventory was completed for the entire Cordillera Blanca using ASTER and SPOT scenes, most of which were from 2001 to 2003 (UGRH, 2010). The majority of glaciers in the Cordillera Blanca are relatively thin and steep, and thus classified as mountain-type in the UGRH inventory. The glaciers are critical for water resources in the region with most draining into the Rio Santa, which originates at Laguna Conococha (4050 m a.s.l.) and eventually flows into the Pacific Ocean approximately 10 km north of
The contributing area above the La Balsa discharge station (~ 5000 km²) constitutes an area that is referred to locally as the Callejón de Huaylas (Figure 2.1). For all analyses in this study, we refer to the Callejón de Huaylas regional watershed as La Balsa. Based on 2007 Peruvian census data, this regional watershed had an estimated 267,000 inhabitants (Mark et al., 2010). Within and adjacent to the Callejón de Huaylas are ten glacierized watersheds, all of which flow to the Rio Santa and have long term records of discharge. These watersheds range from 41 km² to 384 km² in size (Table 2.1).

The Llanganuco watershed is of particular glaciological interest due its large elevation range, large fractional glacier coverage (Table 2.1), variety of glacier type (mountain and debris-covered), and data availability. The contributing area of the watershed is defined by a discharge station below Chinancocha Lake (3850 m). The watershed is adjacent to the highest point in the Cordillera Blanca and all of Peru, Nevado Huascaran (6768 m). The steep west face of the north peak was the site of a catastrophic mass movement, induced by a magnitude 7.8 earthquake in 1970, which left approximately 6,000 people dead (Evans et al., 2009). Because of its history and prominence, the glaciers of Huascaran and its neighbor, Chopicalqui (referred to as the Huascaran-Chopicalqui Massif), have been studied extensively (Georges, 2004; Kaser et al., 1996). Other catastrophic events in other areas of the Cordillera Blanca, such as mass movements and glacier lake outburst floods, have provided additional motivation for studying glacier change (Ames and Francou, 1995).

The defining features of the climate in this region are a pronounced seasonal distinction in precipitation and a lack of seasonal temperature variation. There is a large seasonal distinction in the precipitation totals for the wet (Oct. – April) and dry (May – Sept.) seasons. Based on precipitation records from stations located in the Llanganuco and Querococha watersheds, approximately 90% of precipitation falls during the wet season (Figure 2.2), while discharge is buffered in the dry season by glacier melt. During the wet season, moist air is carried in from the southeast by the Intertropical Convergence Zone (ITCZ), with the Cordillera Blanca serving as a topographical barrier to the rest of the
Callejón de Huaylas and the coastal regions below. In the dry season, trade winds originating from the southeast are dominant (Kaser and Osmaston, 2002). Normally precipitation only falls as snow at higher elevations, usually above 5000 m (Hellström and Mark, 2006). In terms of temperature, the difference between wet season average temperature and dry season average temperature is much smaller than the average diurnal temperature range (Kaser et al., 1990).

The mass balance dynamics of glaciers in this region are different from the majority of glaciers outside of the tropics (Kaser, 1999). Glaciers in the Cordillera Blanca accumulate most of their mass in the wet season (austral summer), but also lose the most mass during this time (Kaser and Osmaston, 2002). Additional glacier melt occurs during the dry season (austral winter), buffering seasonal low flows (Mark and McKenzie, 2007). Only a few glaciers in the region have been studied in detail, namely Artesonraju, Uruashraju, Yanamarey, and Broggi, which was located in the northeastern portion of the Llanganuco watershed. Previous studies have presented traditional mass balance measurements and measurements of terminus retreat for these glaciers (Hastenrath and Ames, 1995; Kaser et al., 1990). Another study made indirect mass balance estimates based primarily on historical monthly discharge (Kaser et al., 2003). These measurements and estimates are useful for inferring patterns of glacier change in the Cordillera Blanca when remote sensing data sets are limited.

2.4 Methods

2.4.1 Data sources and preprocessing

The suitability of satellite imagery for change detection depends on cloud cover, the date of acquisition, and the presence of seasonal snow. For this study we selected nearly cloud-free scenes that were acquired late in the dry season (July to August) when seasonal snow cover is typically at a minimum. We used imagery from five satellites: Landsat 2 Multispectral Scanner (MSS), Landsat 5 TM, Landsat 5 Enhanced Thematic
Mapper (ETM+), ASTER, and IKONOS-2. Table 2.2 lists the spatial resolution and dates of acquisition for the scenes used in this study. Although the Landsat satellites have relatively short repeat cycles (ex: 16 days for TM), the presence of cloud-cover in mountain environments and the seasonal limitation make it particularly difficult to acquire multiple high-quality images for any one year. Furthermore, the rate of glacier change and the errors associated with mapping glacier change using medium-resolution satellites (image co-registration and resolution) only permit change estimates to be made with confidence over time scales of 5-10 years (Hall et al., 2003). Thus, we selected four scenes from the following years for detailed analysis: 1987, 1996, 2004, and 2010.

Landsat scenes downloaded from the United States Geological Survey (USGS) Earth Resources Observation and Science (EROS) Center are already orthorectified and projected. For each year, we acquired a scene from World Reference System (WRS) path 08 and row 66 and WRS path 08 and row 67. Orthorectified and georectified ASTER scenes, including digital elevation models (DEM), were downloaded from the Land Processes Distributed Active Archive Center (LPDAAC). All Landsat and ASTER scenes were projected to the Universal Transverse Mercator (UTM) coordinate system, zone 18 S. The IKONOS-2 scene was projected prior to download. The scene was orthorectified using the Environment for Visualization of Imagery (ENVI) v4.8 with the Shuttle Radar Topography Mission (SRTM) 90 m DEM, provided by the Consultative Group for International Agriculture research – Consortium for Spatial Information (http://srtm.csi.cgiar.org/).

Next, we converted Landsat TM and ETM+ digital numbers (DN) to top-of-atmosphere (TOA) radiance using preprocessing tools in ENVI v.4.8. Atmospheric correction was performed to convert from top-of-atmosphere radiance to surface reflectance. For this we used the atmospheric correction model Second Simulation of a Satellite Signal in the Solar Spectrum (6S; http://6s.ltdri.org/) (Vermote et al., 1997). Table 2.3 shows an example of the parameters used in the atmospheric correction procedure. For each correction, we set the target altitude to 4.5 km above sea level (a.s.l.), which is
approximately the lowest glacier terminus elevation in the Cordillera Blanca. For aerosol optical depth, we used a constant value of 0.1 based on monthly estimates from Multi-angle Imaging SpectroRadiometer (MISR) Monthly Global 0.5 x 0.5 Degree Aerosol Product (MIL3MAE).

Atmospherically-corrected Landsat TM scenes were co-registered to the 18 August 2010 Landsat 5 TM scene if the observed offset was greater than 0.5 pixels. The IKONOS-2 and Landsat ETM+ scenes were co-registered to each other, instead of to the 2010 Landsat 5 TM scene. The IKONOS-2 and ETM+ scenes, both from May 2003, were only used for calibration and validation of the mapping method discussed next.

2.4.2 NDSI threshold calibration and validation

Similar to previous studies in this region (Racoviteanu et al., 2008; Silverio and Jaquet, 2005; UGRH, 2010) we used the NDSI to map glaciers:

\[
\text{NDSI} = \frac{\rho_{VIS} - \rho_{MIR}}{\rho_{VIS} + \rho_{MIR}} \quad \text{(Eq. 2.1)}
\]

where \(\rho_{VIS}\) is the reflectance in the visible part of the electromagnetic spectrum (TM Band 2) and \(\rho_{MIR}\) is the reflectance in the mid-infrared portion of the electromagnetic spectrum (TM Band 5). Compared to other automated glacier-mapping methods, NDSI is particularly advantageous in this region because it is better suited for mapping glacier ice in complex terrain where shadows are common.

Each NDSI image was classified into debris-free glacier and non-glacier zones using a threshold value. The optimal threshold value for debris-free ice was determined by comparing NDSI-generated polygon areas from the 19 May 2003 Landsat ETM+ scene with the area of a hand-digitized polygon from the 11 May 2003 IKONOS-2 scene. For calibration of the NDSI threshold we selected a portion of a glacier terminus outside of Llanganuco where the IKONOS-2 and ETM+ coverage were coincident. We chose to focus on lower elevations of the glacier because these portions of a glacier usually have the NDSI values that are closest to the selected threshold. We varied the threshold from 0.3 to 0.6 in increments of 0.02 and applied each threshold to the ETM+ NDSI image.
We then calculated the percent difference between hand-digitized glacier area in the IKONOS-2 scene and NDSI area from ETM+. We selected the threshold with the smallest absolute percent difference.

To validate this threshold choice, we applied the same threshold to a different, debris-free glacier terminus, also adjacent to the Llanganuco watershed. We created a new subset to use for clipping the Landsat ETM+ NDSI image. We converted NDSI raster-based grid cells to a polygon using ArcGIS v.9.3 and we measured the area of this polygon. Next, we mapped the terminus observed in the IKONOS-2 scene by hand and measured this area. We then compared the difference between the hand-digitized estimate and the NDSI estimate. Unfortunately, there were few additional glacier termini for threshold validation due to cloud cover in the ETM+ scene as well as the limited spatial coverage of this particular IKONOS-2 scene.

To evaluate the effect of atmospheric correction we focused on the Llanganuco watershed. First we masked out non-glacierized and debris-covered glacier areas in ENVI so we could focus only on debris-free ice. We then applied a range of NDSI thresholds to a single TM scene and compared the differences between debris-free glacier area estimates generated using atmospherically-corrected and non-atmospherically-corrected NDSI images. Next, we applied our validated NDSI threshold to two versions of each TM scene, one that had been atmospherically corrected and one that had not been atmospherically corrected. We also included a threshold buffer of 0.1 to illustrate the range of debris-free glacier area estimates that might be observed using different thresholds.

2.4.3 Mapping glaciers in the Cordillera Blanca
The single, validated NDSI threshold was then applied to each Landsat TM scene (1987, 1996, 2004, and 2010). Again we converted reclassified NDSI raster-based grid cells to polygons using ArcGIS v9.3. To remove lower elevation errors, which were usually associated with water bodies and deep shadow, we clipped these polygons to areas above 4000 m a.s.l. Similar to Racoviteanu (2008), we also deleted polygons smaller than 0.01
km² based on our conceptual definition of a glacier. Finally, obvious mapping errors, such as lakes, were removed based on visual inspection of the TM scene being analyzed. Clouds were present over glacierized portions of the Cordillera Blanca in three of the four scenes. However, the extent of the cloud cover was less than about 1% in each case. In cases where a cloud obscured a glacier, we edited the glacier terminus by hand. The terminus was either redrawn to match the extent of the previous year’s outline or points were added in cloud breaks and then interpolated.

We did not apply the NDSI mapping method to the 1975 Landsat 2 MSS scene because the MSS does not record data in the MIR portion of the spectrum that is equivalent to band 5 of Landsat TM and Landsat ETM+. Also, we did not apply the 6S atmospheric correction to this scene because glaciers outlines were created by hand-digitization. The resolution of the scene and saturation over some glaciers made it difficult to map glaciers accurately, especially ice-free areas within the glacier body. Additional attempts to map debris-free glacier ice with supervised classification attempts were unsuccessful. Therefore we do not report area estimates for this scene and instead use it to make some general observations.

Mapping debris-covered glaciers has proven to be very challenging since debris is spectrally similar to nearby terrain. While most studies have relied on manual delineation by hand-digitization (Hall et al., 1992; Racoviteanu et al., 2008), others have attempted to develop semi-automated methods (Bolch and Kamp, 2006; Paul et al., 2004; Taschner and Ranzi, 2002). However, even when a debris-cover mapping method is moderately successful in one location, it is not always transferable to other regions. We tested several different automated debris-cover mapping methods and had little success. Therefore, we hand-digitized debris-covered glaciers using false color composites (bands 5, 4, and 3) of each Landsat TM scene. For each scene, the upper edge of the debris-covered glacier polygons was made coincident with the termini of the debris-free glacier polygons since the termini locations of the debris-free glaciers were different for each scene. The locations of debris-covered glaciers within Llanganuco and the Cordillera Blanca were
confirmed with Google Earth (QuickBird imagery), an IKONOS-2 scene covering most of the Llanganuco watershed (3.2m; acquired on 5 December 2002), and two sets of mosaicked ASTER scenes (15m), each covering nearly the entire Cordillera Blanca.

For each scene, debris-covered glacier polygons were merged with debris-free glacier polygons for an estimate of total glacier area. For our analysis of change in glacier area we focused on the watersheds listed in Table 2.1 as well as the entire Cordillera Blanca. We used the 90 m resolution SRTM DEM along with ArcHydro tools and gauge locations to delineate watershed boundaries. To be consistent with previous studies in the Cordillera Blanca, we estimated the error in mapping glaciers using a one pixel (30 m) buffer method (Congalton, 1991; Racoviteanu et al., 2008; Silverio and Jaquet, 2005). The ±30m buffer was applied to the merged debris-free glacier and debris-covered glacier polygons and the area was measured again for each buffer.

2.5 Results

2.5.1 **NDSI threshold calibration and validation**
We found a threshold of 0.42 to be most accurate for mapping debris-free glacier ice in the Cordillera Blanca. For our calibration subset, the smallest absolute percent difference between NDSI-generated glacier area and glacier area estimated using the high-resolution IKONOS-2 scene resulted from a threshold selection of 0.42 (Figure 2.3). The NDSI-generated polygon for this threshold shows excellent visual agreement with the IKONOS-2 scene (Figure 2.4). We make the assumption that glaciers were located in the same position in each scene and that new snowfall was not significant over the 8 days that separates these two scenes. We chose to validate this threshold in a different subset that is also adjacent to Llanganuco, but has a different aspect. For our validation subset, the debris-free glacier area difference using the 0.42 threshold was approximately 1%. This level of error is acceptable for glacier mapping. There is very good visual agreement between the NDSI validation outline and the observed extent from the IKONOS-2 scene (Figure 2.5).
2.5.2 Effect of atmospheric correction and threshold selection
For the Landsat TM scene from 2010, our threshold choice of 0.42 applied to the Llanganuco watershed uncorrected NDSI yielded an area estimate that was 1.2 km² (5.3%) less than the atmospherically-corrected NDSI (Figure 2.6). We also applied a range of thresholds to both the corrected and uncorrected 2010 NDSI images. The difference between the area estimates from the corrected and uncorrected NDSI images increases as the NDSI threshold is increased.

Next, we compared atmospherically-corrected and atmospherically-uncorrected NDSI debris-free glacier area estimates for each TM scene, again focusing on the Llanganuco watershed (Figure 2.7). For each scene the difference appears to be somewhat systematic. The atmospherically-corrected area estimates are consistently about 5% higher than the atmospherically-uncorrected area estimates. Figure 2.7 also illustrates that a relatively small increase or decrease of the NDSI threshold can have a significant impact on debris-free glacier area estimates. For the corrected image, an applied threshold that is 0.1 too low would result in an average error of 3.7%, while an applied threshold that is 0.1 too high would result in an average error of -4.1%. Similarly, for the uncorrected image an applied threshold that is 0.1 too low would result in an average error of 4.6%, while an applied threshold that is 0.1 too high would result in an error of -5.5%.

2.5.3 Glacier area change across watersheds of the Cordillera Blanca
The resulting glacier outlines derived from the NDSI and hand-digitization of debris-covered glaciers from the years 1987, 1996, 2004, and 2010 are shown superimposed on the Llanganuco watershed (Figure 2.8). These outlines illustrate patterns of area loss within this watershed. The total area change for all sub-watersheds, broken down by time period, is shown in Figure 2.9. For the Llanganuco watershed, the total change in glacier area from 1987 to 2010 was 6.5 km², or 19.5% relative to 1987. For the Rio Santa watershed up to the La Balsa station, the total change in glacier area from 1987 to 2010 was 91.5 km², or 23.2% relative to 1987. Over the entire Cordillera Blanca glaciers lost 161 km², or 25% of their area relative to 1987. Table 2.4 lists the measured glacier areas for each scene and for each sub-watershed.
We next examined how glacier area changed for watersheds as a function of initial glacier extent (Figure 2.10) and initial median glacier elevation derived from the SRTM DEM (Figure 2.11). In general, the sub-watersheds with the most glacier-covered area appear to be most representative of the changes observed for the Callejón de Huaylas and the entire Cordillera Blanca. Watersheds with the lowest median glacier elevation lost the most glacier area from 1987 to 2010. Querococha is an exceptional example of total percent glacier area loss because it is a watershed with a lower maximum elevation and therefore a lower median glacier elevation.

Next, we highlight the changes in total glacier area at three different scales (Figure 2.12). We incorporated data from 11 previous studies in order to illustrate the change in glacier area prior to 1987 and to make comparisons with our area estimates. Mean annual temperature measured at the Querococha meteorological station (see location in Figure 2.1) from 1965 to 1997 is plotted on this figure as well (black circles). From 1965 to 1997, mean annual air temperature increased at a rate of 0.27 °C per decade, which nearly matches the rate observed by Mark and Seltzer (2005) who compiled temperature records from 29 Peruvian stations between 9-11°S. This is the only long-term temperature record from the Cordillera Blanca to which we had access.

The overall pattern of glacier change at different spatial scales appears to be relatively consistent (Figures 2.9 and 2.10). Nearly all watersheds show maximum rates of area loss in the period from 2004 to 2010, an intermediate rate of area loss from 1987 to 1996, a much smaller rate of area loss (or gain in some instances) from 1996 to 2004. However, the overall magnitude of area lost from 1987 to 2010 is more spatially heterogeneous. This is likely a function of median glacier elevation as well as initial glacier coverage from 1987. There does not appear to be a significant difference in total glacier area loss between glaciers that drain to the Pacific Ocean and those draining to the Atlantic. The eastern portion of the Cordillera Blanca (147 km² in 2010) lost 27% of 1987 glacier area, while the western portion of the Cordillera Blanca (336 km² in 2010) lost 25% of 1987 glacier area. We would expect glaciers on the eastern slopes of the Cordillera Blanca to
lose a greater percentage of their area because they are lower on average (Kaser and Georges, 1997). Moving from north to south, there is a non-significant positive trend in total percent glacier area change from 1987 to 2010, meaning that glacier area in watersheds of the southern Cordillera Blanca declined more than glacier area in watersheds of the northern Cordillera Blanca. This trend is likely explained by the higher mean and maximum watershed elevations in the central and northern portions of the Cordillera Blanca.

2.6 Discussion

2.6.1 Atmospheric correction and threshold selection
The major advantage of atmospheric correction is that the same threshold can (theoretically) be used for each scene, assuming that other atmospheric variables, such as smoke or haze, are not introduced from one scene to the next. Relative to uncorrected data, atmospheric correction makes a noticeable difference when estimating glacier area. Figures 2.6 and 2.7 imply that studies mapping debris-free glacier area with an uncorrected NDSI image may be slightly underestimating debris-free glacier area. Figures 2.6 and 7 also illustrate the importance of honing in on the right threshold. Relatively small differences in the choice of threshold can have significant effects on the estimation of debris-free glacier area. For an uncorrected NDSI image, the debris-free glacier area estimate is off by approximately 5% if a 0.1 threshold buffer is applied in either direction. Atmospheric correction thus eliminates analyst error that may result from selecting different thresholds for different scenes because the image is essentially standardized. Atmospheric correction and/or the selection of an accurate threshold would be especially important if the percent change in glacier size from one year to another was the same order of magnitude as the 5% difference observed for the Llanganuco watershed. An example of such a period from this study is the change from 1996 to 2004 when total glacier area in Llanganuco declined by approximately 3%.
Other studies in the Cordillera Blanca have utilized the NDSI to map glacier area. However, two of the three these studies do not discuss atmospheric correction and only one discusses consideration (UGRH, 2010). Silverio and Jaquet (2005) used variable threshold values (0.4 and 0.52) for two TM different scenes. The authors explain that they used a different threshold because reflectance was lower in the August 1996 image, relative to the May 1987 image they selected. Racoviteanu et al. (2008) argued that the atmospheric effects are negligible. They used a threshold value of 0.5 for two SPOT scenes, but did not discuss how the threshold value was chosen. This is a point that is lacking in most studies of glacier area change. The glacier inventory performed by UGRH used satellite images from the ASTER and SPOT satellites with a single threshold of 0.4. The authors do not discuss the selection of a threshold. Retrospectively, it is difficult to estimate errors associated with using atmospherically-uncorrected bands for the NDSI, especially if different sensors are used. Furthermore, it is difficult for us to evaluate the accuracy of estimates from previous studies because of different post-processing steps (glacier size thresholds, manual adjustment for lakes and shadows), as well as differences in the interpretation of debris-covered glacier extent.

2.6.2 Error analysis
There does not appear to be a consensus within the glaciological community on the best method for estimating errors associated with glacier area estimates generated from automated methods (Racoviteanu et al., 2009). Possibly as a result, most studies focusing on glacier area change do not fully quantify the errors associated with glacier mapping (Bhambri and Bolch, 2009). The central problem is that high-resolution “true area” measurements, which could be used for traditional error analyses, are limited, especially for the Cordillera Blanca. The IKONOS-2 imagery used in this study would be useful for measuring “true area.” Unfortunately, we only have access to two scenes, one of which was acquired in December 2002 when snow cover likely obscured actual glacier extent. The other IKONOS-2 scene, acquired in May 2003, was only coincident with a May 2003 Landsat ETM+ scene that had significant cloud cover.
There are several sources of error associated with our estimates of glacier area. First, there is positional error associated with each glacier outline due to imperfect geometric correction. We estimate the positional accuracy of all scenes to be less than 30 m, relative to the 18 August 2010 Landsat 5 TM image. In most cases, the accuracy is better than half a pixel (15 m). An important assumption is that our reference Landsat TM scene had been accurately georeferenced and orthorectified prior to download. It is also important to consider errors associated with topographic illumination. But, as noted by Racoviteanu et al. (2008), these errors should be minimized with the NDSI. As a result, we did not account for misidentification of glacier ice due to shading. There is also uncertainty associated with our choice of threshold. We noted that there was approximately a 1% difference between the debris-free glacier areas of our calibration and validation subsets. In reality, the threshold may be slightly more variable. We were not able to quantify just how much the threshold might vary because of the small size of the glacier area available for calibration and validation.

We also consider “conceptual errors” described by Racoviteanu et al. (2009). We did not define individual glaciers, but we did set a minimum glacier size threshold of 0.01 km² to eliminate snowfields and other small ice bodies that may have been classified as glaciers. Furthermore, our delineation of debris-covered glaciers is interpretive. These delineations were based on previous outlines from the Global Land Ice Measurements from Space (GLIMS) database, the IKONOS-2 scenes, and QuickBird imagery in Google Earth. Our only field-based verification of debris-covered glacier extent is from the Llanganuco watershed.

To lump all of these sources of error together and be consistent with previous studies in the region, we used the GIS buffer method (Congalton, 1991; Racoviteanu et al., 2008; Silverio and Jaquet, 2005), adding and subtracting 30 m from each outline and then calculating the total areas for the resulting glacier polygons. This method tends to overestimate error (Hoffman et al., 2007) and should be considered as more of an absolute range than an error estimate, such as one standard deviation. Estimated error for
each glacier outline produced in this study is shown in Table 2.4. Another option for estimating uncertainty may be to incorporate the uncertainty associated with a variable threshold, as was done in Figure 2.7. In this way, the analyst would be accounting for uncertainty associated with a range of commonly applied thresholds.

2.6.3 Glacier area change over time
We illustrate the change in glacier area over time at different scales using our own estimates and estimates from previous studies (Figure 2.12). Georges (2004) provides a thorough description of glacier extent in the Cordillera Blanca from approximately the Little Ice Age to 1990. Georges notes that glacier retreat rates were not constant over the 20th century, instead the greatest declines were in 1930s and 1940s, with the rate of decline slowing in the 1950s and 1960s, and then increasing again to intermediate retreat rates from 1970 to the end of the 20th century. An analysis of a 1975 Landsat 2 MSS scene suggests a possible advance in the mid-1970s. In the MSS image, the coarse spatial resolution and saturation over some areas make it particularly difficult to map the interior of a glacier body. However, after co-registration the positions of most glacier termini can be mapped reasonably well. Comparisons with outlines from 1970 suggest a slight advance. Other data records also suggest that an increase in glacier area from 1970 to 1975 is plausible and likely. Annual precipitation measured at Llanganuco leading up to 1970 was roughly average, relative to the period from 1953 to 2010. Annual precipitation measured in 1973 and 1975 was more than 250 mm above the long-term average, presumably leading to an increase in glacier area. Furthermore when looking at the long term record of annual temperature from the Querococha station as well as temperature records shown by Racoviteanu et al. (2008), we see that annual temperatures were near their lowest values around 1975. Observations of the Broggi and Uruashraju glaciers (Figure 2.13) (Portocarrero et al., 2008) as well as others (Kaser et al., 1990) suggest that these smaller glaciers were in a state of equilibrium or were even slightly advancing at this time.

From 1975 to 1987 there is little information about glacier area changes available from remote sensing instruments. Glacier behavior in the Cordillera Blanca during this 12 year
period has to be inferred from the long-term reference glaciers already mentioned (Figure 2.13). Based on these observations it appears that glaciers began to retreat again starting in the late 1970s. Similar to Georges (2004), we observed an increased rate of glacier change from 1987 to 1996. During this time the Cordillera Blanca lost 1% of 1987 glacier area per year. Georges (2004) notes a decrease in the rate of glacier area decline after the El Nino in 1997-1998 and that some glaciers even started to advance slightly during this time. These observations are in line with our glacier change estimates from 1996 to 2004. During this time, the Cordillera Blanca lost only 14.6 km², or just 0.3 % of 1996 glacier area per year. This rate of glacier change is about 50% of the average 1970 – 2003 rate reported by Racoviteanu et al. (2008) and also approximately 50% of the average rate of change from 1970 to 2010.

From 2004 to 2010 glaciers lost area at an accelerated rate. Glaciers in the Cordillera Blanca lost 87 km², or 2.5% of 2004 glacier area per year. This is higher than any previously published rate of area loss in the Cordillera Blanca between any two years. Baraer et al. (2012) note an increase in the rate of glacier area decline from 1990 to 2009, relative to the rate observed for the period from 1930 to 2009. They note that the rate increased by approximately 30%. Comparing our estimates from 1987 to 2010 for the La Balsa watershed to the long-term (1930 to 2009) average, we observed a greater than 60% increase in the rate of glacier change per year. The rate of loss from 2004 to 2010 is even greater. During this period, glaciers in the La Balsa watershed lost area at a rate that is nearly 400% greater than the long-term average.

2.7 Summary
This study outlines an effective method for mapping glaciers using atmospherically-corrected satellite imagery and a single threshold value for an automated mapping method. In this study, atmospheric correction effectively standardized each scene of interest, allowing us to choose a single threshold. We validated our threshold choice using high-resolution satellite imagery. This process helps to reduce error associated with
atmospheric effects as well as error associated with the application of variable, unvalidated thresholds. Although we are able to eliminate some error, it is still challenging to estimate the overall error of an individual glacier outline. We used a buffer method which likely overestimates the error associated with our outlines. Future work should develop systematic methods for estimating the error associated with a glacier outline. An additional limitation of this study is that we are not able to robustly quantify the accuracy of our threshold selection. Future work applying the methodology outlined here should calibrate and validate the threshold using multiple glaciers, provided that ground data or high-resolution imagery of some kind is available. In this way, the analyst will be able to determine a single optimal threshold with an associated standard error estimate. This standard error estimate would be one piece of a complete error characterization of a glacier outline generated with an automated method.

Applying a single, validated threshold to atmospherically-corrected Landsat TM scenes, we measured a 25% decrease in glacier area in the Cordillera Blanca from 1987 to 2010. The La Balsa watershed (Callejón de Huaylas), which is home to over a quarter of a million people, lost 23% of glacier area from 1987 to 2010. Watersheds with the least glacier extent and lowest median glacier elevations lost more area than their counterparts over this period of time. The rate of glacier change is not temporally consistent as the decline in glacier area appears to be accelerating based on the most recent estimates from 2004 and 2010. This accelerated rate is a serious cause for concern as communities in this arid environment continue to cope with changing water supplies. Future work focusing on mapping glacier change in the Cordillera Blanca should utilize the NDSI in order to be consistent with prior studies, as well as our own. Furthermore, future studies should explore new methods for accurately mapping debris-covered glaciers in this region as their role in the water balance is not completely understood.

Since on-the-ground mass balance measurements are very limited in the Cordillera Blanca, accurate estimates of glacier area, along with the error associated with these estimates, will continue to be important for understanding how glaciers are responding to
a changing climate. Finally, accurate estimates, like the ones provided here, are essential for hydrologic models which utilize these values to calculate current glacier contributions to streamflow and make predictions about future glacier extent and runoff under different climate scenarios.
2.8 Tables

Table 2.1: Physical characteristics of the study watersheds. Percent glacierized area was derived by Mark and Seltzer (2003) from 1962 aerial photographs. Elevations are from the SRTM DEM.

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<th>Watershed (Station Name)</th>
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<th>Watershed Area km²</th>
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<th>Min. Elevation m.a.s.l.</th>
<th>Max. Elevation m.a.s.l.</th>
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Table 2.2: Resolution and dates of acquisition for each satellite scene used in this study.

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Table 2.3: Example 6S input parameters from the August 2010 Landsat TM scene.

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Table 2.4: Glacier area estimates for each of the sub-watersheds as well as the entire Cordillera Blanca. Error was estimated using the one pixel (30 m) buffer method.

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2.9 Figures

**Figure 2.1**: Location of the study sub-watersheds and the Callejón de Huaylas (referred to here as La Balsa) watershed within the Cordillera Blanca. Glacier extent from August 2010 is also shown.
Figure 2.2: Average monthly precipitation (P) and discharge (Q) measured at Llanganuco (41% glacierized in 1962) and Querococha (6% glacierized in 1962).

Figure 2.3: The NDSI threshold was selected as the value that produced the smallest absolute percent difference between NDSI mapped glacier area and glacier area mapped with a 3.2 m resolution IKONOS-2 image.
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Figure 2.5: Threshold validation of glacier terminus outline derived using Landsat ETM+ with NDSI threshold of 0.42 (a). The NDSI-derived glacier outline from (a) is superimposed on the IKONOS-2 image for comparison (b).
Figure 2.6: Comparison of NDSI-derived glacier areas for Llanganuco derived from the 2010 Landsat TM scene before atmospheric correction (no AC) and after atmospheric correction (AC).

Figure 2.7: Comparison of debris-free glacier areas in Llanganuco derived using atmospherically-corrected (circles) and atmospherically-uncorrected scenes (squares). Each point corresponds to an area measurement derived using the NDSI with a threshold of 0.42. The bars associated with each point correspond to minimum and maximum NDSI threshold values of 0.32 and 0.52, respectively.
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Figure 2.9: Area change in each of the study sub-watersheds and the entire Cordillera Blanca from 1987 to 2010. Area loss for each sub-watershed is shown with a pie chart. White sections of the pie chart represent the total remaining glacier area (2010) while other sections represent area lost between three periods (1987 to 1996, 1996 to 2004, and 2004 to 2010). Black text within the pie chart indicates the total area loss from 1987 to 2010.
**Figure 2.10:** Fraction of 1987 glacier area for each sub-watershed and the Cordillera Blanca. Symbol size for each year is proportional to initial 1987 glacier area.

**Figure 2.11:** Relationship between 1987 median glacier elevation (derived from SRTM DEM) and total percent area change from 1987 to 2010.
Figure 2.12: Glacier area change in the Cordillera Blanca at different scales. Mean annual temperature measured at the Querococha station (black filled circles) is shown in the top graph. Error bars are shown for years when error was estimated.
Figure 2.13: Field observations of the retreat of the Broggi and Uruashraju glaciers from an initial terminus location (Portocarrero et al., 2008).
3 Distributed modeling of runoff in a glacierized basin of the Peruvian Andes: assessing the impact of glacier recession on water resources

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3.1 Abstract

Glacier melt from the Cordillera Blanca, Peru buffers stream flow in the Rio Santa watershed. Glaciers in this mountain range have been losing area rapidly over the last 40 years and many studies predict that these changes will have negative impacts on water availability, especially during the dry season. Previous hydrologic modeling studies have used simplistic empirical or hybrid-type hydrologic models because of data availability limitations in the Cordillera Blanca. We attempt to take the next step forward in understanding the hydrologic processes in this region. Focusing on one watershed of the Cordillera Blanca, we use a physically-based hydrological model: the Distributed Hydrology Soil Vegetation Model (DHSVM) with a newly-coupled dynamic glacier model. The model is run with spatial input files that relate to soil type and depth, vegetation, and topography. We used bias-corrected reanalysis data (MERRA) as our meteorological input and ran the model from 1988 to 2007. Model calibration and validation were performed with static glaciers and with dynamic glaciers. A large number of model runs were required to calibrate the model only somewhat successfully. Modeled and observed glacier changes were in agreement on an annual basis. However, given the current calibration methodology, model complexity, and the lack of validation data, we do not yet have enough confidence in our model to make predictions about future runoff in the Cordillera Blanca. Continued installation of meteorological stations and field
measurements of glacier mass balance will be necessary for this promising methodology to be applied successfully in this region.
3.2 Introduction

Most glaciers around the world are declining in area and volume and this translates to a loss in seasonal water storage for populations that are adjacent to highly glaciated mountain ranges, such as the Rocky Mountains, Himalaya Hindu-Kush, and Andes. On a regional scale, the hydro-societal importance of glaciers varies based on population density, precipitation patterns, and climate (Kaser et al., 2010). Over large areas and with a mixture of different hydrological and climatic forcings, the glacier melt signal is often difficult to separate from other flows. However, on smaller, more local scales the hydrological impact of glaciers is more evident (Nolin et al., 2010; Pelto, 2011). In the Santa River (locally referred to as the Rio Santa) watershed of the Cordillera Blanca (Figure 3.1) this is especially true during the dry season since the glacier melt contribution to discharge is estimated to be approximately 40% (Mark et al., 2005). These vital water reservoirs have been decreasing in size as well. From 1970 to 2010 glaciers in the Cordillera Blanca lost 28% of their area (Georges, 2004) (see Chapter 2). Looking to the future, climate models project that air temperatures will increase faster at higher elevations (Bradley et al., 2006) and many project that these and other associated climatological changes will have a strong, negative impact on glaciers in the Cordillera Blanca (Barnett et al., 2005; Juen et al., 2007).

In general, glacier melt models can be divided into two categories: empirical models and physically-based models. Empirical models require fewer input parameters than physically-based models, but sometimes perform just as well (World Meteorological, 1986). One type of empirical model that is commonly applied is the temperature-index model. These types of models require temperature, evaporation, and precipitation data. They are based on the premise that glacier melt is empirically related to air temperature (Hock, 2003). Within these models, the degree-day factor is usually one of the most important model-tuning parameters. Temperature-index methods are often successful because of the strong relationship between air temperature, longwave radiation, and sensible heat. Ohmura (2001) argued that, on average, incoming longwave radiation and sensible heat comprise approximately 75% of the energy available for melt. This point
was debated by Sicart et al. (2008) who argued that incoming longwave radiation is poorly correlated with air temperature when cloud emissions are the major source. Sicart et al. (2008) examined correlations between air temperature, the surface energy balance, and melt. For the Zongo glacier, located in the Bolivian Andes, they determined that air temperature is a poor index for melting on time steps that are smaller than one year.

Studies utilizing physically-based models are not as common because of the large number of input parameters required to force the model (typically temperature, relative humidity, wind speed, net radiation, and precipitation, at a minimum). Furthermore, it is challenging to maintain stations that measure these parameters in mountain regions and acquire continuous records. In the Cordillera Blanca, data limitations have resulted in studies which do not use completely physically-based models.

Previous studies focusing on glacier melt in the Cordillera Blanca have addressed the vulnerability of the Callejón de Huaylas to changing water supplies. Pouyaud et al. (2005) simulated runoff in four glacierized watersheds that feed into the Rio Santa. The study used an empirical model which was based on the correlation of melt and air temperature taken from the NCEP-NCAR reanalysis at the 500 hPa level. Glaciers were represented as a single class and there was no distinction between the accumulation and ablation areas. In this model, the rate at which the surface melts is based on topographic observations from the Yanamarey glacier, from which the authors derived a coefficient of glacier retreat.

Mark and Seltzer (2003) and Juen et al. (2007) note that the relationship between air temperature and runoff in the Cordillera Blanca is very poor. Juen et al. (2007) argues that temperature index models are thus not applicable. As a result, Juen et al. ran the ITGG-2.0-R on a monthly time step, simulating past and future discharge in the Llanganuco watershed of the Cordillera Blanca from 1953 to 1997. The authors estimated base flow to be approximately 20% of precipitation from the previous month plus a constant value. Moisture-related variables were linearly-derived from precipitation data that was averaged over six stations. Forcing data was derived from meteorological
stations in the Cordillera Blanca and the Cordillera Real (Bolivia). The model was run using different air temperature data: first using data from the Querococha station (75 km to the south) and then using data from the NCEP-NCAR reanalysis at 500 hPa. A key limitation of this study is that glacier area was held constant throughout the model simulations. Glacier-covered area in the Cordillera Blanca declined by approximately 13% from 1970 to 1996 (see Chapter 2). Despite a lack of transient glacier simulations, the authors were still able to accurately reproduce observed discharge in Llanganuco ($r^2 = 0.76$).

Suarez et al. (2008) used a conceptual semi-distributed model (Schaefli et al., 2005) in the highly-glacierized Paron watershed to quantify the contributions of glacier meltwater to streamflow. In this model, glaciers were divided into contributing and non-contributing zones which are defined by the position of the snow line. The model, which required only temperature and precipitation input data, was run on a monthly time step and the glacier extent was held constant. Although Schaefli et al. (2005) used three parallel reservoirs, the authors chose to only use a glacier reservoir for simulations. Considering its simplicity, the model performed reasonably well for the entire Paron watershed, although it tended to overestimate peak discharge.

Chevallier et al. (2011) note the high correlation between monthly NCEP-NCAR reanalysis temperature at the 500 hPa level and monthly discharge measured in the Llanganuco watershed. They make predictions about future discharge based on the relationship between reanalysis temperature and discharge, arguing that slight increases in discharge are to be expected in the next 20 to 50 years. However, the authors acknowledge that this correlation will not be valid after glaciers have retreated significantly, since the relationship between melt and air temperature will have changed.

Condom et al. (2011) modified the Water Evaluation and Planning (WEAP) model and applied it to the Rio Santa watershed. This study was novel because it adopted a semi-distributed approach and modeled the hydrologic impact of decreasing glacier size. Glacier melt was simulated using a degree-month approach that was not properly
justified. Glacier area evolution was simulated successfully for the entire watershed using an area-volume scaling relationship. Nash-Sutcliffe efficiency (NSE) values for the 16 study sub-watersheds ranged from 0.19 to 0.72, with an average of 0.54, indicating that the model performed reasonably well overall but was not consistent. The authors do not provide any quantified predictions about future stream flows in the Rio Santa.

Most recently, Baraer et al. (2012) analyzed historical records of annual discharge from nine sub-watersheds that drain into the Rio Santa in conjunction with a conceptual model of glacier retreat and its influence on discharge. The model uses a simplified water balance approach. Glacier area change was simulated based on satellite observations while volumes were estimated using scaling relationships. The authors found weak correlations between discharge parameters and precipitation totals. They argue that that discharge has already peaked for seven of the nine glacierized watersheds they studied in the Cordillera Blanca.

In contrast to previous work done in the Cordillera Blanca, we simulate runoff including glacier and snow melt, using a physically-based, fully-distributed hydrologic model. Furthermore, we incorporate glacier dynamics to simulate the glacier evolution over the study period. We acknowledge that this approach is still somewhat limited due to data availability, but given the previous work it is the next step forward in understanding the hydrological role that glaciers play in watersheds of Cordillera Blanca. The main objectives of this study are to:

(1) use statistical tests to determine what effects glaciers may have already had on streamflow,

(2) determine the sensitivity of the most important model parameters, and

(3) calibrate and validate the Distributed Hydrology Soil and Vegetation Model (DHSVM) for the Llanganuco sub-watershed using a reanalysis data set from 1988 to 2007. Calibration and validation are based on monthly streamflow data as well as observed glacier extent.
3.3 Study Area

The Cordillera Blanca, located in the central Andes of Peru (8.5°S to 10°S) (Figure 3.1) has the highest concentration of glacier anywhere in the tropics (Kaser et al., 1990). This mountain range lost between 22 and 24% of glacier-covered area between 1970 and 2010 (depending on which 1970 area estimate is used) (see Chapter 2). Based on multi-temporal analysis of Landsat TM scenes, glacier retreat in the Cordillera Blanca has accelerated between 2004 and 2010 (see Chapter 2). Climatological conditions in the Cordillera Blanca are similar to an outer tropical regime. In terms of glacier mass balance, almost all accumulation occurs during the pronounced wet season (Oct. – Apr.). The majority of ablation occurs during this season as well. There is less melt during the dry season (May – Sept.) due to higher rates of sublimation and evaporation coupled with decreased longwave radiation and periods of higher albedo (Juen, 2006; Kaser and Osmaston, 2002).

The source waters of the Rio Santa are derived from snow and glacier melt from many of the high mountains of the Cordillera Blanca. The area draining into the Santa River to the La Balsa station is commonly referred to as the Callejón de Huaylas. This regional watershed has an area of nearly 4800 km² and is home to approximately 267,000 people (Mark et al., 2010). Within the Callejón de Huaylas there are eight glacierized watersheds with long term monthly records of precipitation and discharge. These stations were originally installed by a Peruvian hydroelectric company in the 1950s, but since the early 2000s they have been managed by the Peruvian government and various collaborators.

We chose to focus on one of these watersheds, namely Llanganuco, because of the available data and the fact that previous studies have tested other hydrologic models here. This watershed has average monthly precipitation and discharge measurements that date back to 1953 and 1954, respectively. Starting in 2003, discharge measurements were made at a sub-daily interval using a pressure transducer. Interestingly, an analysis of historical discharge by Baraer et al. (2012) found no trend in dry season discharge over the course of the study period. However, there is a strong relationship between average monthly flows from Llanganuco and the La Balsa station, especially for the lowest flows.
In terms of meteorological instrumentation, a HOBO automated weather station named Llan A was installed in 2004 to measure precipitation, temperature, relative humidity, wind speed, and solar radiation at a sub-hourly interval (Hellström and Mark, 2006). Researchers at Ohio State University installed three additional weather stations in 2006: North Wall, Portachuelo, and Vaqueria. Ten Lascar temperature and humidity data loggers were also installed at various locations throughout the watershed (Figure 3.1).

The watershed ranges from 3850 to 6670 m a.s.l. and has an area of 89 km² based on the Shuttle Radar Topography Mission (SRTM) digital elevation model (DEM) and discharge gauge location (Figure 3.1). The gauging station was installed beneath a large, naturally-dammed lake named Chinancocha. This lake has an estimated depth and volume of 28 m and 1.14 x 10⁷ m³, respectively. The Llanganuco river begins roughly 6 km upstream of the discharge station, at the confluence of the Ancush and Demanda sub-watersheds. The Ancush tributary comes in from the left side, cascading from the debris-covered Kinzl glacier. The Demanda tributary comes in from the right side, travelling a route that is less steep, has more tributaries, and is laden with groundwater storage zones called pampas, which are flat, treeless areas that are thought to be rich in organic matter (Mark and McKenzie, 2007).

Geomorphologically, Llanganuco is a classic U-shaped glacier valley with steep granodiorite walls and thin soils above the valley floor. The valley floor is filled with a variety of lacustrine, colluvial, alluvial, and glaciological deposits. The conceptual schematics presented by Caballero et al. (2002) from the Bolivian Cordillera Real would appear to apply to this region as well in that higher elevation runoff from glacier melt or precipitation takes two primary paths to reach the main channel of the Llanganuco river, which is located in the valley bottom. The first path can be described as slow flow that drains along the face of the steep bedrock walls or through fractures in the intrusive granodiorite. This water next comes in contact with lateral moraines or talus deposits.
which abut the steep valley slopes. The water flows slowly through these deposits and then either re-emerges as a spring or continues beneath the surface until intersecting a channel. The second path is quick flow, or “concentrated” surface runoff, over the thin soils and bedrock of the higher elevations. These paths bypass complete infiltration into the deposits on the valley floor, flowing as surface water directly to the main channel.

A detailed classification of geomorphological features and soil types is lacking for Llanganuco and the rest of the Cordillera Blanca. One of the few detailed studies in Cordillera Blanca, performed by Rodbell (1991), examined soil data from moraines in six high valleys. Most samples from the Cordillera Blanca were classified as either sandy loams or loamy sands and had depths of less than 100 cm. Only two of these samples were taken from Llanganuco. The first site is greater than 110 cm deep, most of which was characterized as loamy sand. The second site is greater than 150 cm depth and has a sandy texture. The approximate percentage of sand, silt, and clay from these two samples is 85%, 10%, and 5%, respectively. Another study by Tremolada et al. (2008) profiled soils along three transects in and adjacent to Llanganuco, ranging from 3710 m.a.s.l. to 4790 m.a.s.l. Ten samples from the A horizon were textured and classified according to the World Reference Base for Soil Resources. The authors note that soils are relatively young and lack cambic horizons. The only soil groups present are Leptosols, Regosols, and Umbrisols.

Medium to high-resolution classifications of land cover are also lacking for the Cordillera Blanca and sub-watersheds. Silverio and Jaquet (2009) used a 2002 Landsat Enhanced Thematic Mapper (ETM+) scene to create a land cover classification for the Huascaran National Park Biosphere Reserve. They used the Normalized Difference Vegetation Index (NDVI) and defined vegetation classes based on relative density and elevation. The authors include common vegetal associations for each class. Byers (2000) discussed landscape change throughout the Huascaran National Park. Byers describes the area below 4800 m as predominantly grassland with remnant polylepis forests. He focused on 5 landscapes with historical photographic records available for comparison, one of which
has a view of Yanapaccha peak from the Pisco basecamp in Llanganuco. Comparing photos from 1939 and 1997, he notes that there had been little change to the Polylepis forests in view. Byers also made measurements of basal area, relative density, relative dominance, and average DBH for two species (Gynoxis and Polylepis) found in a 400 m² plot at 4530 m in Llanganuco. Individual characteristics of four vegetation zones, namely Polylepis forest, Lagunas, Demanda, and Portachuelo Road, were described in detail by Lamas and Perez (1983). The authors provide a thorough inventory of the plant species that are present in each of these zones, but offer few descriptions of the physical characteristics of each species.

3.4 Methods

3.4.1 Trend analysis
We applied the Mann-Kendall trend analysis test (Kendall, 1948; Mann, 1945) to observations of discharge (1954 to 2007) and precipitation (1954 to 2004) measured within the Llanganuco watershed. We also applied this test to temperature data collected from the Querococha watershed from 1965 to 1997. In general, the Mann-Kendall test is used to determine whether the dependent variable increases or decreases with time. It is a non-parametric test which does not require the data to be normally distributed. The sign of the test statistic, $\tau$, indicates the slope of the trend. The slope is significant if the $p$-value is less than the established significance level which is set at 0.05 for this study. Test statistics were calculated for each month and for either the annual total (runoff and precipitation) or annual average (temperature) of each parameter.

3.4.2 Model summary
We use the DHSVM coupled with the University of British Columbia (UBC) glacier dynamics model to simulate past streamflow and glacier extent. The model-coupling process is described by Naz et al. (in prep.). The DHSVM is a physically-based, fully distributed hydrologic model that was originally developed by Wigmosta et al. (1994). The model was designed to represent the effects of soils, vegetation, and complex
topography on hydrologic processes. Each grid cell in the watershed is given a land cover and soil classification. The water and energy balance are solved at each cell for each time step. A 1-D water balance is calculated for each cell based on the effects of vegetation, climate, soil properties, and topography. Vegetation is broken down into two layers: forest canopy overstory and understory. The physical characteristics of these layers are used in calculations of evapotranspiration (Penman-Monteith method) and interception. A two-layer mass- and energy-balance model that incorporates the effects of topography and vegetation on snow is used to simulate snow accumulation and ablation (Andreadis et al., 2009). Unsaturated soil water movement in multiple rooting zones is calculated with Darcy’s Law. Subsurface water is routed in three dimensions as a function of topography and local hydrologic conditions. Surface water flows either through a drainage network created using a DEM or as overland flow in cases where the water table intersects the surface.

At a minimum, the DHSVM requires the following meteorological inputs for each time step: temperature, wind speed, relative humidity, shortwave radiation, longwave radiation, and precipitation. These variables are distributed in the model based on elevation, slope, and aspect. The DHSVM has commonly been applied in mountainous forested watersheds of the Pacific Northwest (Bowling et al., 2000; Storck et al., 1998; Surfleet et al., 2010; Thyer et al., 2004). Other studies have applied the model in glacierized watersheds (Chennault, 2004; Donnell, 2007). However, in these studies glaciers were modeled simply as a static snowpack. The newly coupled version of the DHSVM still uses a two layer snowpack, but now includes a single glacier layer underneath this snowpack. Once the snowpack has disappeared, the glacier becomes exposed and is ablated using the energy balance equations that are used to calculate snow melt. Glacier dynamics and mass balance are modeled on a monthly time step. Mass balance is calculated by the DHSVM and cell to cell ice-fluxes are calculated accordingly (Garry Clarke, pers. comm.).
3.4.3 Spatial input files
The DHSVM requires spatial input files, which are typically created using a Geographic Information Systems (GIS), and a meteorological forcing file. The following spatial files are required for the watershed of interest: DEM, monthly shading files, stream network, land cover type, soil type, soil depth, glacier extent, bed topography, and mass balance. We used the 90 m SRTM DEM (Figure 3.3a) to create monthly shading files, a stream network, and a soil depth map. Efforts to collaborate with other groups and government organizations regarding pre-existing land cover and soil maps have been unsuccessful. As a result, we performed our own land cover and soil classifications.

We mapped different vegetation classes within Llanganuco using an atmospherically corrected NDVI image calculated from bands 3 and 4 of a July 1987 Landsat TM scene. Raw data for bands 3 and 4 were first converted to radiance using Environment for Visualization of Imagery (ENVI) v4.8 preprocessing tools. We then accounted for atmospheric effects by applying the atmospheric correction model Second Simulation of a Satellite Signal in the Solar Spectrum (6S; http://6s.ltdri.org/) (Vermote et al., 1997). Since our minimum model resolution is constrained to 90 m, classes are simplified and generalized. We started by creating three initial land cover classes which were similar to those defined by Silverio and Jaquet (2009). The three initial classes, namely bare, grassland, and shrubland, were based on a stratification of the 1987 NDVI. Since stratification of the NDVI was not representative of our conceptualized classes we appended the three-class raster created in the first step by manually adding additional land cover classes. We first added glacier extent derived from the same Landsat TM scene. A discussion of the methods used to map glacier extent can be found in Chapter 2. We note that debris-covered glaciers are modeled as bare rock since these glaciers have not decreased significantly in size over the course of the study period. Next, we manually added nine lakes which we outlined in the Landsat TM image. Finally, we added a woodland class, which consists mainly of polylepis trees, and a pampa class (Figure 3.3b). Physical characteristics of each land cover (vegetation) class were estimated based on field measurements and literature (see Appendix B.2).
Since there are no high resolution maps of soil type for Llanganuco, we created a simple soil classification that was based on our land cover classification and textural estimates from the 2010 field season (Figure 3.3c). We classified the majority of the soils in the lower portion of the valley as sandy loam. We also created a class for pampas. The soils between the valley floors and recently deglaciated areas are classified as loamy sands. These soils are generally found on steeper slopes and thus are very thin. The higher elevation gravels/bedrock class sits beneath and adjacent to the glaciers in the valley. This class is characterized by thin “soils” which have high infiltration rates but low hydraulic conductivity. Soil depth was estimated using a program that accompanies DHSVM. Soil depth at a particular location is based on the slope at that location as well as the absolute minimum and maximum soil depths estimated for the entire watershed (Figure 3.3d). We estimated the minimum and maximum soil depths for the entire valley to be 0.5 m and 3 m, respectively. Physical characteristics of each soil class were estimated based on field measurements and literature (see Appendix B.3).

Next, we estimated glacier bed topography in order to estimate the ice thickness associated with our initial glacier extent from 1987. This was done using an inversion technique (Garry Clarke, pers. comm.). At a minimum, this method requires the following input parameters: equilibrium line altitude (ELA), mass balance gradient in the accumulation zone, mass balance gradient in the ablation zone, and thinning rate. Our estimates for these parameters are limited since only a few glaciers in the entire Cordillera Blanca have been studied in detail over a period of more than a few years. The terminus retreat of glaciers Broggi and Uruashraju has been measured since 1948. In terms of mass balance, data are very sparse for the Cordillera Blanca. Mass balance measurements were made on glaciers Uruashraju and Yanamarey from 1977 to 1987. Because it is difficult to make direct mass balance measurements on these glaciers (Kaser and Osmaston, 2002), the ELA is normally estimated using other methods. Kaser and Osmaston (2002) estimated an ELA between 5000 and 5250 m a.s.l. in 1970. Juen (2006) estimated an ELA of 5100 m around the year 1990. Mark and Seltzer (2005) used a range of accumulation area ratios (AARs) to estimate an ELA of glaciers in the Guesgue...
watershed. As they note, AARs in the tropics can range from 0.5 to 0.75. Using this range and the hypsometry derived from the 1987 glacier extent and the SRTM DEM, we estimate an ELA range of 5050 – 5250 m a.s.l. (Figure 3.4). Based on these estimates as well as those from previous studies, we used an ELA of 5100 m for estimating bed topography.

More uncertainty is associated with the accumulation and ablation mass balance gradients. We estimated balance gradients using mass balance data from the Yanamarey glacier (Hastenrath and Ames, 1995) collected from 1977 to 1988. It is important to note that this glacier is approximately 80 km to the southeast of Llanganuco and it has a much narrower elevation range (4500 – 5100 m). However, in the absence of other, more complete mass balance measurements, we used these data as a starting point. We estimate an ablation zone balance gradient of 0.025 m/m ice equivalent (i.e.) and an accumulation zone balance gradient of 0.008 m/m (i.e.) from Figure 3.5. We also estimate a thinning rate of 3 ma⁻¹ below ELA and 1 ma⁻¹ above ELA. Again, these estimates are based on the study by Hastenrath and Ames (1995). The modeled ice thickness map is shown in Figure 3.6.

Glacier thickness data for validation of our bed thickness map are also lacking. The first estimates of glacier thickness in the Cordillera Blanca were made by Thompson (1992) using short-pulse radar. These measurements were made on the col of Huascarán in 1983, at Copap in 1983, and at Pucahircra in 1984. Thicknesses on the col of Huascarán (n=6) ranged from 123 m to 192 m in 1980. In Copap, thickness estimates (n=2) range from 197 m to 201 m. On Pucahircra, thickness estimates (n=15) ranged from 159 m to 237 m. Thompson et al. (1995) again used short-pulse radar to survey a coring location. In 1993, two cores located between the twin summits of Huascarán were drilled to bedrock. The two cores were 160 and 166 m long while maximum thickness in the area surveyed was 218 m. The locations of the survey points in both of these studies are not exact (Lonnie Thompson, pers. comm.) but the data can still be used for a rough validation comparison. Glacier thickness data from the year 2009 are also available for the Yanamarey and
Gueshgue glaciers which are located in the southern Cordillera Blanca (Bryan Mark, pers. comm.). The particular transect on Yanamarey has an average depth of approximately 65 m while the transect done on Gueshgue has an average depth of approximately 90 m.

3.4.4 Meteorological input files

The availability of high-temporal meteorological forcing data in the Cordillera Blanca is very limited. High-temporal resolution data is necessary for the DHSVM in order to simulate diurnal hydrological processes, such as snow and glacier melt. There is only one long-term (greater than 20 years) record of mean monthly temperature (Querococha station; Figure 3.1). Monthly precipitation measurements, which date back to 1953 in some cases, are available for a number of watersheds in the Cordillera Blanca, one of which is Llanganuco. Over the past decade, a number of automated weather stations have been installed throughout various watersheds of the Cordillera Blanca with the intent of using the data for mass balance modeling and/or hydrologic modeling. Three of these stations, maintained by Dr. Bryan Mark at Ohio State University and others from UGRH in Peru, are located within Llanganuco. However, none of the meteorological stations are located on a glacier. The station Llan A (elevation 3850 m a.s.l.) began recording data in 2004 and its longest continuous record extends from 2007 to 2010. This station measured temperature, wind speed, relative humidity, incoming shortwave radiation, and precipitation. However, as a result of vandalism, this station failed in 2010 and was removed. Between 2006 and 2012, the station with the most data is Portachuelo (4775 m a.s.l.), which measures temperature, wind speed, relative humidity, incoming shortwave radiation, and precipitation.

In order to simulate discharge and glacier evolution over a longer time period we needed to use a reanalysis data set which has a longer, continuous record, and encompasses all of the necessary meteorological forcing parameters. Reanalyses are retrospective analyses which utilize climatological observations and numerical weather models to create a synthetic record of how the Earth’s weather and climate have changed over multiple decades. The most commonly used reanalysis products include: European Centre for
Medium-Range Weather Forecasts (ECMWF) 40 Year Reanalysis (ERA-40) (Uppala et al., 2005), ECMWF Interim Reanalysis (ERA-Interim) (Dee et al., 2011), Japanese 25-year Reanalysis (JRA-25) (Onogi et al., 2007), National Center for Environmental Prediction / National Center for Atmospheric Research (NCEP/NCAR) Reanalysis I (Kalnay et al., 1996), NCEP Climate Forecast System Reanalysis (CFSR) (Saha et al., 2010), and NASA Modern Era Reanalysis for Research and Applications (MERRA) (Rienecker et al., 2011).

We selected the MERRA reanalysis product from Global Modeling and Assimilation Office (GMAO) and Goddard Earth Sciences Data and Information Services Center (GES DISC) at NASA (http://gmao.gsfc.nasa.gov/research/merra/) because of its relatively fine spatial and temporal resolution. MERRA grid cells measure 0.5° latitude by 0.66° longitude, which equates to approximately 55 km and 70 km, respectively, in the Cordillera Blanca. MERRA is a reanalysis of the satellite record which focuses on historical analyses of the hydrological cycle on a broad range of weather and climate time scales. The grid cell center that is located closest to Llanganuco (Grid Point (GP) 1) has the coordinates -9°S, -77.33°W and an elevation of 3520 m based on surface geopotential height. The next grid point to the south (GP 2) has an elevation (3860 m) that is closer to the elevation of station Llan A (3850 m). In terms of temporal resolution, we acquired hourly MERRA data from 1979 to 2012. Details on how this data set was downloaded can be found in Appendix C.1.

When station data from Llan A were available we compared them to the MERRA data to check for bias. We used all available station data from 2004 to 2010 and extracted MERRA data which matched the availability of the station data. We calculated monthly averages as well as hourly averages for each month for temperature, wind speed, relative humidity, and shortwave radiation. Since station Llan A did not measure longwave radiation, we had to compare MERRA longwave from the Artesonraju glacier station (4850 m). Monthly averages from April 2004 to March 2005 were taken from Juen
(2006) and compared with monthly averages from MERRA for the same time period. Figures showing these comparisons can be found in Appendix C.2.

We used a bias correction methodology which is similar to Berg et al. (2003). We applied a difference-based correction for air temperatures and a ratio-based correction for wind speed, relative humidity, shortwave radiation, and longwave radiation. All corrections, with the exception of longwave, were applied based on hourly averages for each month. This method allows us to correct for bias on annual and diurnal time frames. Longwave data was corrected on a monthly basis since we did not have access to finer temporal resolution data. We applied the bias correction factors derived from the comparison of raw MERRA data and station observations to each hour of each month from 1980 to 2010.

We used a different method to correct raw MERRA precipitation data. Based on an initial analysis of the MERRA precipitation data, it appeared that MERRA overestimated the amount and frequency of precipitation (see Appendix C.2). In fact, the total measured precipitation at station Llan A from 1980 to 2010 was approximately 21.4 m, while the total MERRA precipitation was 88.3 m at GP1 and 68.3 m at GP2. Reichle et al. (2011) note that MERRA is biased high along tropical coastlines of South America. Another issue that they mention, and that we observed as well, is that MERRA precipitation is less intense and is seemingly more of a slow drizzle instead of distinct events.

In order to correct the precipitation data we first used the station precipitation measurements to calculate the average number of days in each month that had precipitation. We then went through each month and ranked the MERRA daily precipitation totals. For each month, if the rank fell below the average number of days with precipitation then precipitation for that day was set to 0 m. We then rescaled the ranked daily totals based on the hourly fraction of daily precipitation from the raw MERRA data. Next we calculated the ratio of measured monthly precipitation to MERRA monthly precipitation totals, adjusted for the average number of days with precipitation, for each month from 1980 to 2010. For each month, we multiplied this ratio
by the hourly, corrected MERRA data. This step helped to ensure that monthly MERRA totals were consistent with historical records of monthly precipitation. Lastly, we noticed that the hourly frequency of precipitation events was still too high, so we applied a precipitation threshold of 0.025 mm. Even after the application of this threshold, the precipitation frequency is still too high relative to measurements made at Llan A, but imposing a greater threshold would result in a significant reduction in total precipitation over the model time period. The 0.025 mm threshold resulted in a loss of only about 350 mm over the 31 year period.

3.4.5 Model calibration and validation

We chose to calibrate the model using available discharge data as well as snow/glacier extent. Our calibration period was from 1988 to 1997. From 1954 to 2003, only monthly discharge data are available. From 2003 to 2008, daily discharge data are available. However, it should be noted that there was a change in instrumentation: in 2003 IRD installed a CHLOE IEL512 pressure transducer. In 2008, a Solinst levelogger was installed. The height difference above the channel bottom between the two sensors is not known. Therefore, since there was no correction performed, data after March 2008 cannot be used. Since we only have access to approximately five years of daily discharge data, we decided to calibrate the model on a monthly time step. Our validation period is from 1998 to 2007. Model performance after calibration and validation runs is assessed using three different criteria: the root mean squared error (RMSE), coefficient of determination ($R^2$), and Nash-Sutcliffe efficiency (Nash and Sutcliffe, 1970).

The RMSE is a measure of overall difference between the model output and measurements and is presented in the same units as the measurements. However, it does not provide a good sense of model predictive skill.

$$RMSE = \sqrt{\frac{\sum_{i=1}^{n}(P_i-\bar{O})^2}{n}} \quad (Eq. \ 3.1)$$
\( P_i \) and \( O_i \) are modeled and observed discharge, respectively, at a time step, \( i \). The coefficient of determination is computed as:

\[
R^2 = \left( \frac{\sum_{i=1}^{n}(O_i - \bar{O})(P_i - \bar{P})}{\sqrt{\sum_{i=1}^{n}(O_i - \bar{O})^2 \sum_{i=1}^{n}(P_i - \bar{P})^2}} \right)^2 \tag{Eq. 3.2},
\]

NSE is computed as:

\[
NSE = 1 - \frac{\sum_{i=1}^{n}(O_i - P_i)^2}{\sum_{i=1}^{n}(O_i - \bar{O})^2}. \tag{Eq. 3.3}
\]

An NSE equal to 1 means that the model perfectly matches the observations. A negative NSE means that the model performs worse than simply using the mean of the observations. A positive NSE means that the model has some degree of predictive skill.

Due to the large number of input parameters required by the model and large uncertainty associated with some of these parameters, we decided to only calibrate first-order model parameters, or those which have the largest effect on modeled streamflow. The first order parameters that we selected for calibration are all related to precipitation and snow or glacier melt. These parameters are temperature lapse rate, precipitation lapse rate, rain/snow threshold, and albedo decay. We argue that other model parameters related to soil and vegetation type (see Appendix B.2 and B.3) are less important than these first-order model parameters. With the exception of the valley bottom, soils are generally very thin in Llanganuco. Similarly, we would not expect the relatively sparse vegetation in the watershed to have a significant effect on discharge.

The standard temperature lapse rate used in most hydrological models is \(-6.5^\circ C \) km\(^{-1}\). Kaser (2001) and Juen et al. (2007) used this lapse rate, while Suarez et al. used different lapse rates for the wet (0.49°C km\(^{-1}\)) and dry (0.54°C km\(^{-1}\)) seasons based on observations between two stations. WorldClim data (Hijmans et al., 2005) covering the model extent also suggest a temperature lapse rate that varies between the wet \((-5.9^\circ C\)
Local observations from the meteorological stations nearest to Llanganuco suggest that the precipitation lapse rate may be very small. Figure 3.7 shows how monthly precipitation totals compare from four stations either within or adjacent to Llanganuco. These stations range in elevation from 2527 m to 4775 m. Based on monthly totals from precipitation stations throughout the Cordillera Blanca, Kaser and Osmaston (2002) and Juen et al. (2007) used a precipitation lapse rate of 0.030 kg m⁻² m⁻¹ and 0.035 mm m⁻¹ month⁻¹, respectively. The imbalance between mean annual runoff (1060 mm) and mean annual precipitation measured at station Llan A (629 mm) from 1954 to 2004 suggests that the precipitation lapse rate must be positive in order for mean annual precipitation to be greater than mean annual runoff. DHSVM calculates precipitation at different elevations using the following equation:

\[ P_z = P_{z_0} \times (1 + (y \times (z - z_0))) \]  

(Eq. 3.4)

where \( P_z \) is precipitation at an elevation above or below the station, \( P_{z_0} \) is the precipitation measured at the station, and \( y \) is the precipitation multiplier. The slope of the line fit through \((z - z_0)\) versus \( P_z \) is equivalent to the precipitation lapse rate. We tested a range of precipitation multipliers from 0 (0 mm m⁻¹ year⁻¹) to 0.01 (6.3 mm m⁻¹ year⁻¹).

In many modeling studies the rain/snow threshold is assumed to be 0°C. However, Dai (2008) showed that the phase transition over land can occur between -2°C and 4°C. Dai also notes significant pressure dependence at very high elevation (surface pressure < 750 hPa) because precipitation falls faster in thinner air. Based on the SRTM DEM, the greatest fraction of watershed area is located at approximately 4900 m, which converts to a surface pressure of approximately 550 hPa. The DHSVM requires specification of the maximum temperature which snow will fall at as well as the minimum temperature that rain will fall at. Based on the figures in Dai (2008), we selected a phase transition range
of 1°C to 3°C. The DHSVM partitions rain and snow linearly using this defined range, meaning that 50% of precipitation will occur as snow at 2°C.

The standard snow albedo decay equation used by the DHSVM is described by Andreadis et al. (2009) and Thyer et al. (2004), and is based on work by Laramie and Schaake (1972). The equation assumes that snow albedo decays exponentially with age, but that snow albedo decreases more quickly if the snow is melting ($T_s = 0^\circ$C). The albedo decay functions derived by Laramie and Schaake (1972) have the following form:

\[
\begin{align*}
\text{if } T_s < 0, & \quad \alpha_S = \alpha_0 A_a (N)^B_a \\
\text{if } T_s = 0, & \quad \alpha_S = \alpha_0 A_m (N)^B_m
\end{align*}
\] (Eq. 3.5) (Eq. 3.6)

where $\alpha_S$ is the snow surface albedo for the current time step, $\alpha_0$ is the albedo of fresh snow, $N$ is the number of days since last snowfall, and $A$ and $B$ are the decay coefficients, where the subscripts $a$ and $m$ correspond to melt and accumulation. Based on measurements by Storck (2000), $A_a$ is 0.92 and $B_a$ is 0.58, while $A_m$ is 0.72 and $B_m$ is 0.46. Fresh snow albedo typically ranges from 0.8 to 0.97, with an average value of 0.84, while the minimum albedo of firn is estimated to be 0.43 (Paterson, 1994).

In terms of glacier albedo, we initially used a constant albedo of 0.5 based on measurements from Artesonraju glacier, located in the neighboring Paron watershed (Juen, 2006). In later model runs, we tested an albedo parameterization from Oerlemans (1992) in which albedo varies as a function of elevation above or below the ELA:

\[
\alpha_g = a + b \times \arctan \left( \frac{h-E+c}{d} \right)
\] (Eq. 3.7)

where $\alpha_g$ is the albedo of the glacier at a certain elevation, $h$, $E$ is the ELA, and $a$, $b$, $c$, and $d$ are coefficients. Our parameterization of glacier albedo with elevation is

\[
\alpha_g = 0.38 + 0.05 \times \arctan \left( \frac{h-5100+300}{200} \right)
\] (Eq. 3.8).
For glaciers in our watershed, this equation produces albedos which range from approximately 0.34 to 0.45 depending on the elevation.

In our initial calibration runs, the dynamic glacier model was turned off. We used the July 1987 glacier outlines derived from a Landsat TM scene as our initial glacier extent and an ice thickness of 1000 m for all glaciers. We used unrealistically large thicknesses to ensure that the glaciers would not disappear at the margins while the model was running. In later runs, we turned the glacier dynamics model on and used the initial ice extent and thickness modeled by the bed topography inversion technique. We turned on the glacier dynamics model in order to illustrate the impact of glacier change on streamflow.

3.5 Results and Discussion

3.5.1 Trend analysis
To determine what effect changes in glacier size have had on past discharge, we first examined the trend in monthly discharge from 1954 to 2007. The results of the Mann-Kendall test (Table 3.1) suggest that there has been no significant increase or decrease in Llanganuco discharge for any month of the year. Furthermore, there has not been a significant increase in total annual runoff. The lack of a trend for any month is perplexing since this watershed has declined in glacier area by 8.1 km², or 23%, since 1962 (Mark and Seltzer, 2003). We might expect precipitation to outweigh the effects of changes in glacier size during the wet season. However, we would certainly expect to see a trend in dry season discharge if glaciers have been the primary contributor to streamflow. However, when applying the Mann-Kendall test no trend is observed for any month during the dry season. To remove extreme discharge values which may have been obscuring a significant trend, we examined the interquartile range of discharge from 1954 to 2007 (Figure 3.8). A linear regression of these discharge values with time still does not show any significant change in discharge with time.
We also applied the Mann-Kendall test to monthly precipitation totals measured in Llanganuco from 1954 to 2004 (Table 3.2). While nine months had negative trends, only one of these trends (September) was significant. The trend observed in September is due to the decrease in variance which is observed starting in 1987. The frequency of small monthly rainfall totals (less than 5 mm) in September increased noticeably during this time.

Lastly, we applied the Mann-Kendall test to average monthly temperatures recorded at the Querococha station from 1965 to 1997 (Table 3.3). All trends were positive in sign, but only 5 were significant. It is interesting to note that all of the significant trends are associated with wet season months. The trend in annual average temperature is highly significant (p=0.001). Racoviteanu et al. (2008) performed the same statistical test on temperature records from 3 stations in the Cordillera Blanca and found a significant, positive trend as well.

Previous work (Baraer et al., 2012) has shown that discharge from watersheds in the Cordillera Blanca that are experiencing reductions in glacier area usually increases initially and then begins to decline steadily. Analyzing Llanganuco discharge records from 1954 to 2007, we observed no trend in discharge despite a significant loss in glacier area. Possible explanations for the lack of a trend are evapotranspiration (ET) and sublimation. In terms of transpiration, it is not likely that vegetation cover has increased significantly over the past 25 years (Byers, 2000). Field exploration of recently deglaciated areas in Llanganuco confirms this postulate. However, evaporation rates potentially could have increased with the formation of additional lakes and higher air temperatures. Similar to evaporation, water that is sublimated is transferred to the atmosphere instead of the stream. Sublimation is a significant component of the surface energy balance of glaciers in the tropics since it reduces the amount of energy available for melt (Wagnon et al., 1999; Winkler et al., 2009). This process is more prevalent in the dry season when relative humidity is at a minimum. Juen (2006) estimated an annual sublimation total of approximately 150 mm w.e. based on measurements collected from
the Paron watershed. Winkler et al. (2009) used a lysimeter to estimate sublimation on the tongue of glacier Artesonraju during the 2005 dry season. Daily sublimation totals ranged from 1 – 5 mm d\(^{-1}\).

### 3.5.2 Model calibration and validation using static glaciers

The current version of DHSVM does not have a routine for sensitivity analysis or parameter optimization. Therefore, although the values of the parameters we report correspond with our best model runs (as determined by the NSE criteria), the values of the input parameters we report may not necessarily be optimal. The precipitation lapse rate which resulted in the best set of efficiency criteria was 0.32 mm m\(^{-1}\) month\(^{-1}\) \((y = 0.0005)\). The model was highly sensitive to this parameter, especially during the wet season. We did not change the coefficients of the snow albedo decay functions, but we did vary the value for fresh snow albedo and minimum snow/firn albedo. We obtained the best results using a fresh snow albedo of 0.85 and a minimum snow/firn albedo of 0.45. This choice of minimum snow/firn albedo also led to a relatively smooth albedo transition from melting snow to exposed ice. We found that the model is highly sensitive to the value of fresh snow albedo (Figure 3.9.a). Changes as small as 0.01 can have noticeable effects on the modeled discharge. The model is also sensitive to changes in glacier albedo (Figure 3.9.b), although the discharge associated with a variable glacier albedo is tied to the presence and frequency of snow cover.

The results of our best calibration run using static glaciers are shown in Figure 3.10. We note that this particular run used the variable glacier albedo instead of a constant value. First, we see that the model performed relatively well during the calibration period. The NSE is 0.5 while the RMSE is 0.78. Mass balance estimates appear to be reasonable when compared to model output from the ITGG-2.0 mass balance model (Juen, 2006). Unfortunately, ITGG annual mass balance was only calculated until 1994, so we have very few years for comparison. The comparison between modeled and observed average monthly discharge for the entire calibration period shows that simulated streamflow is too low during the first half of the year and too high from September to December. Positively-biased flows from September to December could be explained by solar
radiation values at this time of the year. August and September have the highest average incoming shortwave radiation after the model accounts for the effects of topography and atmospheric transmissivity. Increased values of incoming shortwave radiation correspond with the highest monthly values of glacier melt.

The results of the validation run associated with this calibration run are shown in Figure 3.11. The NSE decreased to 0.39 while the RMSE increased to 0.85. For the combined calibration and validation runs, it is surprising that using the static glacier extent works as well as it does. We might expect the simulated discharge to be positively biased during later years when the glacier extent should be reduced. This suggests that the glacier meltwater component of discharge may be less important than the snow melt component. Based on our model setup, this does indeed appear to be the case. Over the course of the year, basin-averaged values of snow melt are approximately twice as great as basin-averaged values of glacier melt (Figure 3.12.a). This ratio increases during the wet season and approaches 1 during the dry season. We also see that average monthly snow melt is highly correlated with average monthly modeled discharge, while glacier melt and discharge are not highly correlated (Figure 3.12.b).

3.5.3 Model calibration and validation using dynamic glaciers

Next, we switched on the dynamic glacier model and calibrated the DHSVM using observed average monthly discharge. The initial (1988) area of the modeled glaciers is 26.1 km², while the debris-free glacier area observed using a satellite scene from 1987 is 30.3 km². This difference in glacier extent can be explained by errors in the SRTM DEM or by unrepresentative input parameters used for bed topography estimation. Figure 3.6 shows the initial measured and modeled glacier extent. The northern portion of the watershed is the only area where there is a noticeable difference in extent. Glaciers in this portion of the watershed have slightly lower termini elevations and likely lower ELAs since their maximum elevations are also lower. The western orientation of these glaciers and their proximity to the wetter eastern slopes could explain why the glaciers extend to lower elevations. Furthermore, in the Cordillera Blanca, diurnal patterns of cloudiness
and precipitation allow glaciers to extend to lower elevations on western slopes (Kaser and Georges, 1997).

From our calibration runs with dynamic glaciers, the precipitation lapse rate which resulted in the best set of efficiency criteria was 0.32 mm m\(^{-1}\) month\(^{-1}\) \((v = 0.0005)\). We found a fresh snow albedo value of 0.82 and a minimum snow/firn albedo value of 0.45 produced the best model results in terms of discharge. In terms of glacier albedo, we calibrated the model using the variable glacier albedo parameterization (Eq. 8).

The results of our best calibration run using dynamic glaciers are shown in Figure 3.13. Again, the model performed relatively well during the calibration period. The NSE is 0.45 while the RMSE is 0.81. The \(R^2\) for the monthly averages was 0.82 which is an improvement on the \(R^2\) values from calibration and validation runs with the static glacier extent. The monthly and annual specific mass balance estimates appear to be reasonable, but it is difficult to discuss their accuracy without actual glacier mass balance data from this area.

The results of the validation run associated with this calibration run are shown in Figure 3.14. The NSE decreased to 0.23 while the RMSE increased to 0.96. Based on monthly averages (Figure 3.14.e), modeled streamflow is too low for most of the year. However, this version of the model does appear to do a better job of simulating the average seasonal variability in discharge. Modeled specific mass balance is very low during the validation period. Less than a quarter of all months have a positive specific mass balance. Over the course of the year, basin-averaged values of snow melt are again approximately twice as great as basin-averaged values of glacier melt (Figure 3.15.a). This ratio is highest during the wet season and approaches unity during the dry season. Again, we also note that average monthly snow melt is highly correlated with average monthly discharge \((R^2 = 0.81)\), while glacier melt and discharge are not highly correlated (Figure 3.15.b). Therefore, our model suggests that average monthly discharge reflects monthly patterns of snow melt more closely than glacier melt.
Next, we compared the modeled and observed glacier extents. We mapped debris-free glacier area for the years 1987, 1996, 2004, and 2010. Unfortunately it is difficult to resolve glacier extent changes at a finer resolution using satellite imagery. Based on monthly area estimates, the modeled glaciers appear to be very dynamic, showing large changes in area within a single year (Figure 3.16.a). To better compare the modeled glacier area with our observations, we calculated the annual average modeled glacier area for each year. Since the initial modeled and observed areas are not quite the same, we compared relative changes, showing the fraction of initial area that remained for each year with an observation (Figure 3.17.b). The relative modeled area changes generally match the trend that we observed.

3.5.4 Predictive validity of the model and next steps
Ideally, we would use the knowledge and confidence gained from model calibration and validation to make predictions about future runoff in the Cordillera Blanca under different climate scenarios, similar to Juen et al. (2007). However, the calibration results produced by the model thus far, as well as a lack of rigorous validation data, inhibit this next step. Previous hydrologic modeling studies in the Cordillera Blanca have been able to simulate observed discharge just as well, if not more accurately, using models that are more simplistic (see section 3.2). Since the model is physically based, we would expect it to have the potential to be more accurate than other empirical and hybrid-type models that have been used before. The major problem with complex, physically-based models, like DHSVM, is that all of the free parameters create many more degrees of freedom than can be constrained by available calibration data. Simply put, even if there is a set of parameters that produces optimal fits between observed and modeled discharge and glacier extent, it would be very difficult to have confidence in the model given the lack of validation data and the amount of uncertainty associated with the parameters.

Thus far, we have demonstrated the functionality and potential of the model. The next steps involve ensuring that we are getting the right answer for the right reasons (Kirchner, 2006) and that we are doing so in an efficient manner. For example, we were able to simulate observed discharge somewhat successfully using a certain set of parameters, but
given the large number of input parameters, it is likely that there are other sets of parameters which might yield the same result. This is commonly referred to as the problem of model equifinality (Beven, 1993).

Our current method of calibrating the DHSVM is not efficient and does not help us to address the problem of equifinality. Typically we start calibration with a physically “ideal” set of input parameters and then change one to two parameters for each calibration run. Since we are running the model on an hourly time step for at least five years at a time, the calibration process is very time intensive. Furthermore, the process requires the modeler to have some knowledge of parameter sensitivity. While individual sensitivity runs are less time intensive, it is still difficult to understand model sensitivity if multiple parameters are being changed. One option for moving forward would be the application of parameter optimization software, such as PEST (http://www.pesthomepage.org/) (Doherty and Hunt, 2010). PEST is a software package for parameter estimation and uncertainty analysis of complex environmental and other computer models. Implementing this software is an option but this would require significant changes to DHSVM source code and configuration file.

The glacier model will also need to be refined. This model was originally designed to be run at large scales, such as entire mountain ranges like the Canadian Rockies. Typically the model is run at 200-m resolution on an annual time step. The initial glacier extent might be improved by making adjustments to initial parameters required for the estimation of bed topography. Another option for improving the initial glacier extent is to “grow” glaciers to steady state using a very long spin-up period and a spatially-distributed mass balance field. However, this method is problematic because we lack detailed, continuous records of glacier mass balance in this region. Furthermore, if we were attempting to match the initial (1987) glacier extent, we would not be growing glaciers to a steady state, but rather a transient one. Kaser (1990) and Georges (2004) state that glaciers probably have not been at steady state since the period 1970-1975.
Finally, we acknowledge that we have made a number of assumptions in order to simplify the calibration process. For example, we assume that modeled discharge is not very sensitive to changes in soil and vegetation parameters. One of our most important assumptions may have been modeling debris-covered glaciers in Llanganuco (~2.8 km²) as bare rock. We modeled them this way because we did not interpret significant changes in debris-covered glacier extent from 1987 to 2010. But because we do not have accurate estimates of debris-covered glacier retreat or thinning rates this assumption could be invalid. Understanding the hydrologic role of debris-covered glaciers in this region will be a challenge as we move forward.

3.6 Conclusions and future work
We were able to calibrate and validate a newly coupled glacio-hyrdrolgical model, the DHSVM, for a glacierized watershed of the Peruvian Andes. We gained valuable insights regarding model sensitivity. Based on calibration runs and the first-order parameter adjustments, the model is highly sensitive to changes in precipitation and albedo. Although we were able to calibrate and validate the model with some success, the predictive skill is relatively modest based on established efficiency criteria and a lack of rigorous validation data. Nash-Sutcliffe efficiencies for all calibration and validation runs while greater than zero, were less than or equal to 0.5. During our model period, there are no published values of glacier mass balance for Llanganuco. A lack of mass balance data makes it very difficult to determine if the dynamic glacier model is performing well. Since our predictive skill is not strong and we cannot currently estimate the total uncertainty associated with the model, it would be unwise to use the current model to make hydrologic projections associated with different climate scenarios.

Despite our predictive limitations, the need for further study in this area is evident. Previous studies have made projections about future hydrologic regimes with reduced glacier extent, but none have taken glacier dynamics into account. As meteorological and glaciological time series increase in length and accuracy, physically-based glacio-
hydrologic models will provide important diagnostic and prognostic information for water resource managers. The lack of field measurements and meteorological data have inhibited us from providing useful water resource projections. We advocate for the continued field monitoring of glacier mass balance in the Cordillera Blanca as well the maintenance of existing and installation of additional micro-meteorological stations. Furthermore, there is a need for better understanding of the spatio-temporal variations in precipitation and albedo. Albedo decay curves for snow and ice should be established for this region. Lastly, future studies should investigate the behavior of debris-covered glaciers since their hydrologic role is not yet understood in this region.
3.7 Tables

**Table 3.1:** Results of the Mann-Kendall test applied to Llanganuco monthly discharge measurements from 1954 to 2007. Annual runoff is the total runoff for each calendar year. Significant trends are in bold font.

<table>
<thead>
<tr>
<th>Month</th>
<th>T</th>
<th>p-value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Jan</td>
<td>0.01</td>
<td>0.911</td>
</tr>
<tr>
<td>Feb</td>
<td>0.11</td>
<td>0.257</td>
</tr>
<tr>
<td>Mar</td>
<td>0.05</td>
<td>0.612</td>
</tr>
<tr>
<td>Apr</td>
<td>0.08</td>
<td>0.395</td>
</tr>
<tr>
<td>May</td>
<td>0.03</td>
<td>0.720</td>
</tr>
<tr>
<td>Jun</td>
<td>-0.01</td>
<td>0.946</td>
</tr>
<tr>
<td>Jul</td>
<td>0.00</td>
<td>0.982</td>
</tr>
<tr>
<td>Aug</td>
<td>-0.02</td>
<td>0.805</td>
</tr>
<tr>
<td>Sep</td>
<td>0.03</td>
<td>0.754</td>
</tr>
<tr>
<td>Oct</td>
<td>0.11</td>
<td>0.233</td>
</tr>
<tr>
<td>Nov</td>
<td>0.07</td>
<td>0.469</td>
</tr>
<tr>
<td>Dec</td>
<td>0.09</td>
<td>0.317</td>
</tr>
<tr>
<td>Annual Runoff</td>
<td><strong>0.10</strong></td>
<td><strong>0.31</strong></td>
</tr>
</tbody>
</table>

**Table 3.2:** Results of the Mann-Kendall test applied to Llanganuco monthly precipitation measurements from 1954 to 2004. Annual precipitation is the total precipitation for each calendar year. Significant trends are in bold font.

<table>
<thead>
<tr>
<th>Month</th>
<th>T</th>
<th>p-value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Jan</td>
<td>-0.12</td>
<td>0.214</td>
</tr>
<tr>
<td>Feb</td>
<td>-0.11</td>
<td>0.255</td>
</tr>
<tr>
<td>Mar</td>
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<td>0.871</td>
</tr>
<tr>
<td>Apr</td>
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<td>0.581</td>
</tr>
<tr>
<td>May</td>
<td>-0.03</td>
<td>0.782</td>
</tr>
<tr>
<td>Jun</td>
<td>0.04</td>
<td>0.701</td>
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<td>Jul</td>
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<td>0.281</td>
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<tr>
<td>Aug</td>
<td>-0.08</td>
<td>0.476</td>
</tr>
<tr>
<td>Sep</td>
<td><strong>-0.26</strong></td>
<td><strong>0.010</strong></td>
</tr>
<tr>
<td>Oct</td>
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<tr>
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<td>0.407</td>
</tr>
<tr>
<td>Dec</td>
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<td>0.214</td>
</tr>
<tr>
<td>Annual Precip.</td>
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<td><strong>0.61</strong></td>
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Table 3.3: Results of the Mann-Kendall test applied to Querococha monthly temperature measurements from 1965 to 1997. Significant trends are in bold font.

<table>
<thead>
<tr>
<th>Month</th>
<th>( \tau )</th>
<th>p-value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Jan</td>
<td>0.33</td>
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<td>Feb</td>
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<td>0.009</td>
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<td>Mar</td>
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<td>Apr</td>
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<tr>
<td>May</td>
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<tr>
<td>Jun</td>
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</tr>
<tr>
<td>Jul</td>
<td>0.23</td>
<td>0.067</td>
</tr>
<tr>
<td>Aug</td>
<td>0.18</td>
<td>0.152</td>
</tr>
<tr>
<td>Sep</td>
<td>0.13</td>
<td>0.316</td>
</tr>
<tr>
<td>Oct</td>
<td>0.16</td>
<td>0.219</td>
</tr>
<tr>
<td>Nov</td>
<td>0.17</td>
<td>0.171</td>
</tr>
<tr>
<td>Dec</td>
<td>0.31</td>
<td>0.015</td>
</tr>
<tr>
<td>Mean Annual Temp.</td>
<td><strong>0.41</strong></td>
<td><strong>0.001</strong></td>
</tr>
</tbody>
</table>
3.8 Figures

**Figure 3.1:** Map of the Llanganuco watershed and location in the Cordillera Blanca, Peru. Glacier extent is shown for 1987 and 2010 (see Chapter 2). Weather station, Lascar data logger, and stream pressure transducer locations indicated.
Figure 3.2: Llanganuco average monthly discharge plotted versus La Balsa average monthly discharge from 1954 to 2008.
Figure 3.3: Spatial input files used in the model: DEM and stream network (a), land cover classes (b), soil type (c), and soil depth (d).
Figure 3.4: Hypsometry of debris-free glaciers in the Cordillera Blanca (no fill, solid black outline) and Llanganuco watershed (grey fill) derived from 1987 glacier extent and the SRTM DEM. Boxes above the histogram indicate the approximate range of ELAs for the Cordillera Blanca and Llanganuco, assuming an AAR of 0.5 to 0.75.
Figure 3.5: Modified from (Hastenrath and Ames, 1995). Vertical net balance profiles of ice equivalent for the period 1977-1988, in meters. Squares correspond to negative balances in the ablation area while filled circles correspond to positive balances in the accumulation area. Trend lines are fit through both groups in order to estimate the vertical gradients in the accumulation and ablation areas.
Figure 3.6: Modeled ice thickness map from 1988 created using the bed topography inversion technique. Ice thicknesses are in m. Observed glacier extent mapped from a 1987 Landsat TM scene is also shown.
Figure 3.7: Monthly precipitation totals from four stations: Yungay, Llanganuco, Portachuelo, and Vaqueria. See Figure 1 for station locations.

Figure 3.8: The interquartile range of July discharge measurements from 1954 to 2007 (open circles) and the change in glacier area since 1962 to 2010.
Figure 3.9: Initial 5 year calibration runs showing model sensitivity to fresh snow albedo (a) and constant glacier albedo (b).
Figure 3.10: Model calibration results for run 311, using static glacier extent. (a) Monthly streamflow calibration with Nash-Sutcliffe efficiency and RMSE. (b) Monthly specific mass balance estimates. (c) DHSVM annual specific mass balance estimates compared with modeled estimates from Juen (2006). (d) Observed versus simulated discharge for each month with an $R^2$ value. (e) Monthly average discharge and standard deviations for modeled and observed discharge. The $R^2$ value corresponds to the relationship between these monthly averages.
Figure 3.11: Model validation results for run 311, using static glacier extent. (a) Monthly streamflow validation with Nash-Sutcliffe efficiency and RMSE. (b) Monthly specific mass balance estimates. (c) DHSVM annual specific mass balance estimates. (d) Observed versus simulated discharge for each month with an $R^2$ value. (e) Monthly average discharge and standard deviations for modeled and observed discharge. The $R^2$ value corresponds to the relationship between these monthly averages.
Figure 3.12: Monthly, basin average snow and glacier melt shown with basin average albedo using static glaciers (a). The relationship between snow melt and discharge (blue circles) as well as glacier melt and discharge (black squares) is shown in (b).
Figure 3.13: Model calibration results for run 339, using dynamic glaciers. (a) Monthly streamflow calibration with Nash-Sutcliffe efficiency and RMSE. (b) Monthly specific mass balance estimates. (c) DHSVM annual specific mass balance estimates compared with modeled estimates from Juen (2006). (d) Observed versus simulated discharge for each month with an $R^2$ value. (e) Monthly average discharge and standard deviations for modeled and observed discharge. The $R^2$ value corresponds to the relationship between these monthly averages.
Figure 3.14: Model validation results for run 339, using dynamic glaciers. (a) Monthly streamflow validation with Nash-Sutcliffe efficiency and RMSE. (b) Monthly specific mass balance estimates. (c) DHSVM annual specific mass balance estimates. (d) Observed versus simulated discharge for each month with an R^2 value. (e) Monthly average discharge and standard deviations for modeled and observed discharge. The R^2 value corresponds to the relationship between these monthly averages.
Figure 3.15: Monthly, basin average snow and glacier melt shown with basin average albedo using dynamic glaciers (a). The relationship between snow melt and discharge (blue circles) as well as glacier melt and discharge (black squares) is shown in (b).
Figure 3.16: Monthly modeled glacier area from 1988 to 2007 (a). The fraction of initial annually-average modeled area and initial observed area are shown in (b).
4 Isotopic variation during the dry season in a glacierized watershed of the Peruvian Andes

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4.1 Abstract
The stable isotopes of water (δ¹⁸O and δ²H) are useful for distinguishing between water sources. These isotopes fractionate systematically and are usually inversely correlated with elevation. Elevation of the source water is normally a very good predictor of an isotopic value provided that phase changes have not occurred after falling as precipitation. We measured water isotopes of 30 stream and groundwater samples collected throughout the Llanganuco watershed of the Cordillera Blanca, Peru during the dry (winter) season in 2011 to characterize the spatial variation in water isotopes within the basin, and the isotopic differences between glacier melt and ground water. We found a weak trend with median subwatershed elevation for δ¹⁸O and δ²H. The slope, or the isotopic lapse rate, of this trend (-2.1‰ per km for δ¹⁸O) is slightly more positive than the world average mountain isotopic lapse rate (-2.8‰ per km for δ¹⁸O). A stronger relationship was found between δ¹⁸O (and δ²H) and percent glacier cover suggesting that percent glacier cover may be masking or dampening the elevation effect, by preferentially releasing high elevation water and through mixing processes associated with glacier melt. Lastly, we used a simple, two-component mixing model to estimate relative contributions of glacier meltwater and groundwater during a point in time from the 2011 dry season. We estimate that the surface water leaving the watershed is composed of approximately three-fourths meltwater and one-fourth groundwater.
4.2 Introduction
The stable isotopes of water, $\delta^{18}O$ and $\delta^2H$, are useful tools for distinguishing between different source waters in a watershed because they fractionate systematically as a result of phase changes. Isotopic values of meteoric waters vary inversely with elevation as a result of Rayleigh distillation. This phenomenon, referred to as the altitude effect (Dansgaard, 1964), has been observed in mountain ranges around the world (Poage and Chamberlain, 2001). Using this knowledge, previous studies have used $\delta^{18}O$ and $\delta^2H$ to characterize different source waters, including old (or pre-event) water (Sklash and Farvolden, 1979), water from different elevation zones (Brooks et al., 2012), as well as both groundwater and glacier meltwater (Mark and McKenzie, 2007; Nolin et al., 2010). Hydrograph separation of stream water using stable isotopes is based on mass conservation and the fact that these tracers are conservative, provided that a phase change does not occur after water falls as precipitation (Kendall and McDonnell, 1998).

A number of studies (Baraer et al., 2009; Cable et al., 2011) have used stable isotopes to separate contributions from snow/glacier melt and groundwater. This task is critical for understanding the hydrologic implications of snow and glacier decline. In this study we focus on a heavily glacierized watershed located in the Cordillera Blanca of the Peruvian Andes where previous work using the stable isotopes of water has suggested an increasing contribution of glacier meltwater (Mark et al., 2010; Mark and McKenzie, 2007). In the Cordillera Blanca, many questions remain about the timing and source of groundwater recharge. Although glacier melt undoubtedly makes up a significant part of seasonal and annual flows (Mark et al., 2005), the availability of groundwater recharged by wet season precipitation could potentially reduce the severity of projected hydrologic scenarios associated with a decline in snow- and glacier-covered area. Therefore, it is important to understand the various sources of water in this area and how these sources vary in space and time.

The first goal of this study was to examine the spatial variation associated with the stable isotopes of water. Previous studies have shown that elevation is a key variable for predicting how the stable isotopes of water vary in space. Given that the watershed is also
heavily glacierized, we examined the relationship with percent glacier cover. We also examined the persistence of these trends at larger scales in the Cordillera Blanca. Our second goal was to use a simple, two-component isotopic mixing model to determine relative contributions of groundwater and glacier meltwater for a point in time during the 2011 dry season.

4.3 Study area
The Llanganuco watershed (centered at -9.05° S, -77.61° W) is situated on the western slopes of the Cordillera Blanca, Peru (Figure 4.1) and drains into the Rio Santa. The watershed ranges from 3850 to 6670 m a.s.l. and has an area of 89 km² based on the Shuttle Radar Topography Mission (SRTM) digital elevation model (DEM) and discharge gauge location. As of August 2010, 30% of the watershed was covered by glaciers (see Chapter 2). Approximately 90% of annual precipitation falls during the wet season (October to April). For this reason, previous tracer studies have neglected dry season inputs from precipitation. To our knowledge, no studies focusing on the Cordillera Blanca have documented the isotopic variation of precipitation throughout the year. During the wet season, moist air is carried in from the southeast by the Intertropical Convergence Zone (ITCZ), with the Cordillera Blanca serving as a topographical barrier to the rest of the Rio Santa watershed (Callejón de Huaylas) and the coastal regions below. In the dry season, trade winds originating from the southeast are dominant (Kaser and Osmaston, 2002). Dry season streamflow in Llanganuco and the rest of the Cajellón de Huaylas is buffered by glacier melt water (Mark and McKenzie, 2007; Mark et al., 2005) and groundwater (Baraer et al., 2009). A comprehensive description of the study area can be found in the second chapter of this thesis.
4.4 Methods

4.4.1 Field sampling
We used a synoptic sampling approach in which water samples from a large number of sites are collected in a relatively short period of time (Baraer et al., 2009; Mark and McKenzie, 2007; Mark et al., 2005; Mark and Seltzer, 2003). Water samples were collected in the Llanganuco watershed during a two week period of the dry season (July 2011). Samples for stable isotopes ($\delta^{18}$O, $\delta^{2}$H) were collected in new High-density polyethylene (HDPE) 30 mL Nalgene bottles. Before filling the bottle, it was triple rinsed at the sample point using the sample water. If the water level was too low, a HDPE syringe was used to extract sample water. Water samples were filled completely and caps were wrapped to prevent evaporation. After sample collection, water samples were subsequently stored in a dark, cool location until analysis.

While in the field we attempted to acquire samples from different source waters, similar to an approach by Nolin et al. (2010). Glacier melt samples were deemed to be those that were in close proximity to a glacier and/or draining directly from a glacier. Groundwater samples were acquired where springs emerged from the ground. At a few groundwater sampling locations, springs were identified using an infrared camera in the early hours of the morning. Finally, main channel water sample sites were sites in the valley bottom that were thought to be a mixture of glacier meltwater and groundwater. GPS points were taken at each sample site. The locations of the Llanganuco sample sites are shown in Figure 4.2.

4.4.2 Laboratory analysis
All isotope data are expressed in terms of $\delta$ values in units of ‰ with $\delta$ calculated as,

$$\delta = \left( \frac{R_{\text{sample}}}{R_{\text{standard}}} - 1 \right) \quad (\text{Eq. 4.1})$$

where, $R$ is the ratio of deuterium to hydrogen atoms or $^{18}$O to $^{16}$O atoms of the sample and the standard is Vienna-Standard Mean Ocean Water (V-SMOW). We first ran samples on the Thermo Delta Plus XL Mass Spectrometer using CO$_2$ equilibration.
technique in the College of Earth, Ocean, and Atmospheric Sciences at Oregon State University. This machine is only set up to measure $\delta^{18}$O. Our isotope analysis accuracy was $-0.02 \pm 0.04\%$ (mean $\pm$ standard deviation of standards) for $\delta^{18}$O based on the repeated measured difference of three standards from known values. Based on duplicate runs, our overall sample precision was $0.04\%$ for $\delta^{18}$O.

Water samples were also analyzed with a LGR LWIA (Los Gatos Research, Mountain View, CA) by the Institute for Water and Watersheds Collaboratory (Oregon State University, Corvallis, Oregon) in order to obtain values both for $\delta^2$H and $\delta^{18}$O. Our isotope analysis accuracy was $0.00 \pm 0.05\%$ and $0.00 \pm 0.23\%$ (mean $\pm$ standard deviation of standards) for $\delta^{18}$O and $\delta^2$H, respectively, based on the repeated measured difference of three standards from known values. Based on duplicate and (in some cases) triplicate runs, our overall sample precision was $0.11\%$ for $\delta^{18}$O and $0.58\%$ for $\delta^2$H.

We also incorporated previous isotopic data from Dr. Bryan Mark at Ohio State University that were acquired from 2004 to 2011. For each year, samples were collected at the same location and on approximately the same date (early July). Values of $\delta^{18}$O and $\delta^2$H were measured with a mass spectrometer (Finnigan MAT Delta Plus coupled to a HDO water equilibrator) in the Ice Core Paleoclimatology Lab at the Byrd Polar Research Center at The Ohio State University. These results are reported with an accuracy of $\pm0.2\%$ for $\delta^{18}$O and $\pm2\%$ for $\delta^2$H.

4.4.3 Examining spatial patterns

We plotted sample locations in ArcGIS v. 9.3. To create a representative stream network we used the 30-m resolution ASTER Global Digital Elevation Map (GDEM), instead of the 90-m SRTM DEM, because of its higher spatial resolution. We used ArcHydro tools to delineate watersheds for each sampling point, except groundwater springs. The DEM was also used to extract the elevations of sample points. However, since we were sampling flowing water that was likely a mixture of water from higher elevations, it was more appropriate to associate a given sample with the median elevation of a subwatershed. Initially we focused on samples collected from the Llanganuco watershed.
We only calculated the median elevations for subwatersheds associated with main channel or meltwater points since the area that contributes to a spring is more difficult to define. We also calculated the percent glacier cover for each subwatershed based on an August 2010 glacier outline (see Chapter 2). We used linear regression to estimate the amount of isotopic variance explained by median watershed elevation and percent glacier cover. Regression analyses were only performed for meltwater and main channel samples. Some data points were excluded from regression analyses. An explanation of these exclusions can be found in the Results and Discussion section.

We also performed a multiple linear regression using median elevation and percent glacier cover as predictors for the observed $\delta^{18}$O and $\delta^2$H data. We used a first order model of the following form:

$$Y_i = \beta_0 + \beta_1 X_{i1} + \beta_2 X_{i2} + \epsilon_i \quad (\text{Eq. 4.2})$$

where, $Y_i$ is the isotopic values of $i$th location, and $X_{i1}$ and $X_{i2}$ are the values of the two independent variables for that location. The parameters of the model are $\beta_0$, $\beta_1$, and $\beta_2$, and the error term is $\epsilon_i$. We performed a partial F-test to determine whether adding a second independent variable significantly improves the prediction of $Y$, granted that either $X_1$ or $X_2$ is already included in the model.

Next, we shifted our focus to a larger scale. Using the isotopic values from synoptic samples collected by Mark and McKenzie (2007), we examined the same relationships described above. When performing linear regression using these data, we only used samples collected from 2004 and 2005 since these dates are closest to our glacier outlines from August 2004. The locations of the synoptic sampling sites are shown in Figure 4.1.

4.4.4 Examining temporal trends

We also examined trends in isotopic values and how they relate to annual precipitation to look for evidence of a spatial/temporal pattern called the amount effect, that is when heavy isotopes are the first to be precipitated, resulting in a moisture source that becomes more and more depleted as more rain falls (Kendall and McDonnell, 1998). We plotted
annual precipitation totals (see Chapter 3) versus isotopic values of annually-collected
dry season water samples (B. Mark, pers. comm.) to determine the effect that antecedent
precipitation had on the isotopic value of stream water.

### 4.4.5 Isotopic mixing model

The relative contributions of glacier meltwater and groundwater were estimated using a
simple isotopic mixing model defined by the following equation:

\[
\frac{Q_{\text{glacier}}}{Q_{\text{MC}}} = \frac{(\delta_{MC} - \delta_{GW})_{\text{glacier}}}{(\delta_{MC} - \delta_{GW})_{GW}} \times 100 \tag{Eq. 4.3}
\]

where, \(Q\) is discharge and the subscripts \(MC, GW,\) and glacier represent water from the
main channel, groundwater, and glacier melt, respectively, and \(\delta\) represents the isotopic
ratio using either \(\delta^{18}O\) or \(\delta^2H\).

On watershed to sub-watershed scales, this method is based on samples that are assumed
to be representative of the source waters. Glaciers are a heterogeneous amalgamation of
many years of precipitation that also varies isotopically as a function of temperature,
amount, and antecedent seasonal precipitation. \(\delta^{18}O\) values measured from ice cores
collected from the accumulation area of a glacier within the Llanganuco watershed range
from -14‰ to -25‰ (Thompson et al., 1995). As a result of this variation, it is more
appropriate to use isotopic values obtained from meltwater streams than individual
samples of melted ice. Guided by the assumption that the ablation area of these glaciers is
well mixed, we expected that all glacier melt samples would have relatively similar
isotopic values.

Groundwater is recharged by precipitation as well as glacier melt. Isotopic values of
precipitation vary between the wet season and dry season (Mark and Seltzer, 2003). The
lower isotopic values of rainfall during the dry season coupled with the assumption that
springs are likely recharged by more local, lower elevation precipitation leads us to
expect groundwater isotopic values that are more positive than glacier meltwater values
which represent more high elevation precipitation.
An important requirement for the mixing model is that the isotopic values from the different sample groups are distinct, or statistically different, from one another. To determine if the groundwater and meltwater groups were statistically different from one another, we performed a $t$-test, assuming unequal variance. We performed this test twice for each isotope. First, we performed the test using all of the samples in each group. For the second round of testing, we excluded certain samples. We excluded the meltwater samples 8 and 9 because these samples were taken from the same channel as 7, but further downstream (Figure 4.2). Similarly, we excluded meltwater samples 2 and 3 to reduce the weight of the samples from the northernmost part of the watershed. Also, we did not include groundwater samples 20 and 21 because their points of emergence from the ground could not be located and thus it was not possible to confirm that they originated from a groundwater source. These samples were collected from two separate small flows that drained down the face of a road cut.

After excluding these points, we calculated the average group values to be used in the mixing model. We calculated the basin-wide groundwater average excluding the samples mentioned in the previous paragraph. We assumed that the isotopic value of all groundwater entering the main channel averaged to this estimated mean. For the meltwater group we calculated a weighted average based on the subwatershed area that is defined by the location of each sampling point. The meltwater and groundwater averages were used to calculate the fraction of glacier meltwater in the lowest portion of the watershed, which is represented by main channel samples 29 and 30.

### 4.5 Results and discussion

#### 4.5.1 Instrument comparison

For our samples from the Llanganuco watershed, we compared $\delta^{18}O$ measured with the LGR and the mass spectrometer (Figure 4.3, Table 4.1). The values from the two machines are in very good agreement. For this comparison, the mean difference between the LGR and mass spectrometer was -0.09‰, with a range of 0.03‰ to -0.26‰. Given
the agreement between the δ¹⁸O values from the two machines as well as the added utility of δ²H, we decided to use the LGR data for all further analyses.

4.5.2 Local meteoric water line

We plotted our measured values of δ¹⁸O and δ²H from Llanganuco in dual isotope space (Figure 4.4) to visualize their position relative to the global meteoric water line (GMWL), as described by the equation (Craig, 1961):

\[ δ²H = 8 \ δ¹⁸O + 10 \ (\text{Eq. 4.4}) \]

The GMWL slope value of 8 is nearly the same as the ratio of equilibrium fractionation factors for H and O isotopes at 25-30 °C. The GMWL y-intercept of this equation is referred to as the deuterium excess (d-excess). Our samples, as well as the Llanganuco samples collected by Bryan Mark from 2004 to 2011, produce local meteoric water lines (LMWL) that have lower slopes than the GMWL. The calculated slopes are 6.6 (this study) and 7.1 (B. Mark, pers. comm.). These slopes are in agreement with values for fractionation relationships for ¹⁸O and ²H in ice-water reactions that predict slopes between 6.2 and 7.4 (Clark and Fritz, 1997), indicating that the melt process has likely influenced these values.

Mark and McKenzie (2007) collected water from tributaries draining from both the Cordillera Blanca and Cordillera Negra. They calculated a slightly lower local meteoric water line slope of 5.2. To show the annual variation of local meteoric water line slopes, we plotted the isotopic values from the synoptic samples from 2004 to 2007 (Figure 4.5). Plotting these samples in dual isotope space yields local meteoric water lines that have slopes ranging from 0 to 5.8. Using samples collected from the Querococha watershed which is only 3% glacierized, Mark and Seltzer (2003) calculated a local meteoric water line slope of 8.3. Although these samples were collected during different years, it seems likely that the slope contrast between Llanganuco (30% glacierized) and Querococha (3% glacierized) is explained by the difference in percent glacier cover.
4.5.3 Examining spatial patterns
In order to calculate isotopic lapse rates, we plotted the Llanganuco $\delta^{18}$O and $\delta^{2}$H values of our samples versus median subwatershed elevation (Figure 4.6). Four Llanganuco samples, all from the highest elevation sites in the northernmost portion of the watershed (numbers 1 – 4), were not used in the regression analyses. Three of the sites (1, 2, and 3) have very small contributing areas. Mark and Seltzer (2003) showed that the interannual isotopic variability of small subwatersheds was much greater than that of larger subwatersheds. Furthermore, it is difficult to delineate small sub-watersheds accurately from a 30m DEM and then calculate median elevation and percent glacier cover based on those delineations. When excluding these samples we also considered the fact that all of the sites (including 4) were sampled approximately a week after the majority of the other samples. Precipitation records from the nearest meteorological station indicate that precipitation fell between the first and second sampling periods. These events could have introduced additional isotopic variation into the system. Alternatively, the measured isotopic values could actually be considered to be representative of their respective subwatersheds. This would suggest that the spatial isotopic variability we observed is due to more complex patterns of precipitation rather than the typical altitude effect. For example, precipitation in these locations may be coming from storms originating from a different location than those that provide precipitation to the rest of the basin.

For the Llanganuco samples, the average isotopic lapse rate for $\delta^{18}$O was -2.1‰ per km while the average isotopic lapse rate for $\delta^{2}$H was -14.6‰ per km. Poage and Chamberlain (2001) compiled isotopic data from 68 previous studies to illustrate the altitude effect for different mountain ranges of the world. Focusing on $\delta^{18}$O values from samples taken below 5000 m, they calculated an average isotopic lapse rate of -2.8‰ per km. For samples collected from below 5000 m in Central and South America, they calculated an average isotopic lapse rate of -2.7‰ per km. Poage and Chamberlain excluded samples from elevations greater than 5000 m in average calculations because there was more scatter in the data, possibly as a result of snow post depositional processes (Niewodniczanski et al., 1981) or the incorporation of another moisture source
from the upper troposphere (Holdsworth et al., 1991). For δ¹⁸O the average isotopic lapse rate from studies that collected samples above 5000 m was -4.1‰ km.

In terms of lapse rates measured near our study site, Mark and McKenzie (2007) observed an isotopic lapse rate of -7‰ per km for δ¹⁸O from groundwater springs in the unglacierized Cordillera Negra, located to the west of the Cordillera Blanca. This lapse rate approximately agrees with the lapse rate of -6‰ per km for δ¹⁸O derived from snow samples collected along the northern ridge of our study basin between Nevado Pisco Oeste and N. Huandoy Este (Niewodniczanski et al., 1981). However, it is important to note that Niewodniczanski et al. excluded three sample points when calculating this lapse rate. The three samples had the lowest elevations of the data set and were taken while moving down the glacier. The authors note that these three samples show an inverse gradient which could be related to post depositional processes. These published gradients are considerably steeper than our estimate and those of Poage and Chamberlain (2001).

Water samples collected from Llanganuco have a less steep isotopic lapse rate because they receive water from glacier melt that we would expect to have a relatively similar value in the dry season because the melt zone is likely at similar elevations throughout the basin regardless of subwatershed elevation. This assumes that the melting portion of the glacier is relatively well-mixed and stores substantially more water than it intercepts from rain and snow (Cable et al., 2011). However, we note that there are studies from other nearby locations that suggest that our calculated lapse rate is reasonable. For example, based on precipitation samples collected along transects in Bolivia, Gonfiantini et al. (2001) report isotopic lapse rates for δ¹⁸O that range from -1.4‰ per km to -2.4‰ per km, depending on whether samples were collected in the wet or dry season.

We also plotted δ¹⁸O and δ²H values of our samples versus percent glacier cover and performed linear regression (Figure 4.7). For δ¹⁸O, the regression indicates that δ¹⁸O decreases by -0.25‰ per 10% increase in glacier cover. For δ²H, the regression indicates that δ²H decreases by -2.1‰ per 10% increase in glacier cover.
We performed multiple linear regression to determine if the combination of the two variables, median subwatershed elevation and percent glacier cover, explained more of the isotopic variance than the individual variables. For median elevation versus δ\textsuperscript{18}O the adjusted R\textsuperscript{2} was 0.57, which was lower than the adjusted R\textsuperscript{2} for percent glacier cover versus δ\textsuperscript{18}O, which was 0.63. The adjusted R\textsuperscript{2} from the multiple linear regression of median elevation and percent glacier cover versus δ\textsuperscript{18}O was 0.61. Therefore, incorporating both independent variables does not increase the amount of isotopic variance explained. It is important to note that median subwatershed elevation and percent glacier cover are strongly positively correlated (Figure 4.8); not surprisingly higher elevation watersheds have a greater proportion of glacial cover.

We computed the partial F statistic to determine if the addition of one independent variable significantly improves the model. Results showed that neither β\textsubscript{1} (percent glacier cover) nor β\textsubscript{2} (median subwatershed elevation) had slopes that were statistically different than 0 when the other variable was considered first. Therefore, the addition of either median elevation or percent glacier cover does not significantly improve the prediction of δ\textsuperscript{18}O as compared to one variable models. Though there was a clear relationship between the dependent variable, δ\textsuperscript{18}O, and the independent variables, the individual tests on the regression coefficients leads to the conclusion that they are equal to zero because of the multicollinearity among the independent variables. While both variables were similar in their ability to explain the variation in δ\textsuperscript{18}O, percent glacier cover was the stronger variable.

For median elevation versus δ\textsuperscript{2}H the adjusted R\textsuperscript{2} was 0.40, which was lower than the adjusted R\textsuperscript{2} for percent glacier cover versus δ\textsuperscript{3}H, which equaled 0.69. The adjusted R\textsuperscript{2} from the multiple linear regression of median elevation and percent glacier cover versus δ\textsuperscript{2}H was 0.70. In this case, incorporating both independent variables only slightly increases the amount of isotopic variance explained. From the partial F statistic test, we found that the slope of the fitted plane, β\textsubscript{1} (percent glacier cover), was statistically different from 0, while slope β\textsubscript{2} (median elevation) was not statistically different than 0.
This means that addition of percent glacier cover significantly improves the prediction of $\delta^{2}H$ over median elevation alone, while the addition of median elevation does not significantly improve the prediction of $\delta^{2}H$ using percent glacier alone. Taken together, these results suggest that glacier cover of a subwatershed is a better predictor of $\delta^{18}O$ and $\delta^{2}H$ than median subwatershed elevation.

In this particular watershed it appears that the glacier signal alters the inverse trend that we might normally expect to observe with elevation. However, given the absence of precipitation samples and isotopic measurements at a range of elevations in the Cordillera Blanca, as well as the complex topography, we do not know exactly how elevation influences precipitation within this watershed. Nevertheless, we feel that the process of glacier melting was altering the expected elevation relationship for several reasons. First, our LMWL slope was indicative of melt water processes rather than following the meteoric water line. Second, our lapse rate was lower than expected, which could be from melt occurring at similar elevations throughout the watershed. Finally, percent glacier coverage was slightly better at predicting isotopic variation between subwatersheds than was median subwatershed elevation.

Our observations from the Llanganuco watershed led us to wonder if the isotopic trends we observed were persistent at larger scales in the Cordillera Blanca. To check for persistence we plotted the dry season isotopic values (Table 4.2) from the synoptic watersheds (Figure 4.1) versus median watershed elevation (Figure 4.9) and percent glacier cover from 2004 (Figure 4.10). Again, we see that the relationship between the isotopic value and percent glacier cover is stronger than the relationship between the isotopic value and median watershed elevation. It is important to note that percent glacier cover does not explain as much of the isotopic variance, relative to the amount of variance explained by percent glacier cover in Llanganuco alone. With the exception of the deuterium slope from 2004 (Figure 4.10.b.), the slopes of the regression of percent glacier cover and isotopic value are similar to the regression slopes generated with the samples from Llanganuco.
4.5.4 Examining temporal trends
Based on the samples collected by Bryan Mark and precipitation measurements for Llanganuco from the same period, we see some evidence for the amount effect (Figure 4.7). The amount effect is typically discussed in reference to a single precipitation event. However, another study in the tropics has shown that seasonal isotopic variations can be associated with the amount of precipitation (Gonfiantini et al., 2001). Analyzing isotopic measurements of precipitation from La Paz, Bolivia, Mark and Seltzer (2003) showed that the most negative isotopic values of precipitation fell during the wet season, which is temporally similar to the wet season of the Cordillera Blanca. We discuss the amount effect in terms of annual precipitation measured prior to the isotope sampling date because we only have isotopic values from one date every year. A linear regression of annual precipitation versus δ\(^{18}\)O shows a negative trend that is not quite statistically significant (p=0.08). The same regression done using δ\(^{2}\)H shows a negative trend that is not significant (p=0.33). These observations provide weak evidence for the amount effect in this region. Mark et al. (2010) argue that the primary reason for the observed isotopic depletion of synoptic water samples collected from 2004 to 2008 was an increase in the contribution of glacier meltwater. Our observations suggest that the amount effect may be another important process to consider in conjunction with potentially increasing glacier meltwater contributions.

4.5.5 Groundwater versus meltwater
The isotopic variation of the different sample groups is depicted as a boxplot in Figure 4.12. For both δ\(^{18}\)O and δ\(^{2}\)H, the meltwater group has the largest interquartile range, while the groundwater group has the largest absolute range. For δ\(^{18}\)O, a t-test shows that the means of the meltwater and groundwater sampling groups are not quite statistically different from one another (p = 0.07). For δ\(^{2}\)H, a t-test shows that the means of the meltwater and groundwater sampling groups are not statistically different from one another (p = 0.11). Therefore when all of the samples for each group are included, the two groups cannot be considered to be isotopically distinct from each other.
Next, we applied the same $t$-test to the sample subsets for each group (see section 4.4.5). We argue that these subsets are more representative of the entire watershed because the groups are not weighted as heavily by areas that had a higher sampling density. When examining the $\delta^{18}$O values from the sample subset, a $t$-test shows that the means of the meltwater and groundwater sampling groups are statistically different from one another ($p = 0.05$). When examining the $\delta^2$H values from the sample subset, a $t$-test shows that the means of the meltwater and groundwater sampling groups are not quite statistically different from one another ($p = 0.08$). For both isotopes, the p-value decreased when using the sample group subsets. Both tests now show that the difference in the mean isotopic values of the two groups is nearly statistically significant at the 5% significance level.

Finally, we use the area-weighted average of the meltwater samples subset (n=5) and the basin-wide average of the groundwater samples subset (n=10) to estimate the fraction of glacier meltwater that is present in the lower portion of the Llanganuco watershed (Table 4.3). The area-weighted average of the meltwater samples subset is $-16.3\%$ for $\delta^{18}$O and $-118\%$ for $\delta^2$H. The basin-wide average of the groundwater samples subset is $-14.7\%$ for $\delta^{18}$O and $-108\%$ for $\delta^2$H. Applying the two-component mixing model using the average value of the two sites in the lowest portion of the watershed (29 and 30), we estimate the percent meltwater contribution to be 75% based on $\delta^{18}$O and 77% based on $\delta^2$H during this dry season. We calculated the standard errors associated with these estimates (Table 4.3) based on equations from Phillips and Gregg (2001). The error estimates have very large bounds due to the small sample size and large standard deviation of the meltwater group samples. Clearly, a more detailed isotopic characterization of these sources is needed to accurately use this mixing model approach, but this exercise indicates the potential for such an approach.

The samples collected in this study and the isotopic ranges observed for each sample group shed some light on the primary sources of groundwater recharge. Based on data from the International Atomic Energy Agency (IAEA;
http://www.univie.ac.at/cartography/project/wiser/) it is likely that precipitation falling during the dry season has isotopic values that are considerably more positive than isotopic values of samples collected during the wet season (Figure 4.13). Assuming that seasonal isotopic variation is similar in the Cordillera Blanca, the majority of groundwater should thus be recharged by wet season precipitation and/or snow and glacier melt. Unfortunately, the samples collected for this study cannot provide information about the magnitude of groundwater recharge from various sources. Additional hydrochemistry sampling could provide more information about this process.

In order to estimate the relative contributions of precipitation and glacier melt to recharge, one would have to collect isotopic measurements of precipitation and snow/glacier melt for each month. It would then be possible to calculate an annual mean isotopic value of precipitation for the watershed by weighting the precipitation measurements by monthly precipitation totals. A long-term weighted average isotopic value of precipitation should be very similar to the basin-wide isotopic value for groundwater. If we were then able to assume a relatively constant isotopic value for glacier melt and if this value was significantly different from the average isotopic value of precipitation, we could use a two component mixing model to estimate the magnitude of groundwater recharge from the two primary sources, snow/glacier melt and wet season precipitation.

4.6 Conclusions and future work
We examined isotopic variation with elevation and percent glacier cover in detail for one watershed of the Cordillera Blanca, Peru. We then examined the same relationships at a larger scale. Understanding contributions of water from various sources is important for this region since melt from shrinking glaciers is an especially important component of dry season flows. Based on simple and multiple linear regression using our samples from Llanganuco, we determined that percent glacier cover is a better predictor of isotopic values than median subwatershed elevation. While we fully acknowledge the importance
of elevation on isotopic variation of precipitation, glaciers may skew this relationship by both preferentially storing high elevation precipitation, and by the complex melt processes and their effects on isotopes that occur as glaciers melt. Using median subwatershed elevations derived from a DEM, we calculated an average isotopic lapse rate of -0.21‰ per km for δ¹⁸O and -14.6‰ per km for δ²H based on samples from Llanganuco.

We also discussed temporal trends in δ¹⁸O and δ²H based on samples collected from 2004 to 2011. More data collection and analyses are needed to address the knowledge gaps associated with temporal isotopic variability in this region. The samples collected for this study were merely a snapshot in time. Future studies should examine both diurnal and seasonal isotopic variations as well as the spatial variance. It is important to understand seasonal variations if the ultimate goal is to quantify the annual amount of water that comes from snow and ice melt. In order to understand the variability of the different sample groups and how they are tied together, sampling of precipitation, streamflow, and groundwater should take place each month, similar to the approach by Mark and Seltzer (2003). Along these lines, one of the most pressing questions at the moment is: What is the primary source of groundwater recharge? If a large portion of the dry season groundwater is recharged by wet season precipitation, then dry season water supply concerns may not be as dire as predicted. On the other hand, if only a small portion of dry season groundwater is recharged by wet season precipitation then many of the dry season water supply concerns will be well-warranted, provided that glaciers in this region continue to decrease in size at the current rate.

It was beyond the scope of this study to determine the timing and magnitude of groundwater recharge. Instead, the primary goal of this study was to estimate the fraction of dry season streamflow that was derived from glacier meltwater and groundwater. We used a simple, two-component mixing model to estimate the relative contribution of glacier meltwater at the watershed outlet. We estimate that approximately three-fourths of the surface water leaving the basin at the time of sampling was sourced from glacier melt.
We acknowledge the simplicity of our mixing model and advocate for studies which have the resources to sample stream water continuously in order to understand diurnal and monthly variations in the contributions from meltwater, groundwater, and precipitation. Future studies should continue to use stable isotopes for mixing models, but should also incorporate additional hydrochemical measurements similar to Mark et al. (2005) and Baraer et al. (2009). These additional measurements will be important for identifying groundwater resonance times and sources of recharge.
### 4.7 Tables

**Table 4.1:** Group, location, and isotopic values for each site located in Llanganuco. The sample numbers are also located on Figure 4.2.

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<th>Sample Number</th>
<th>Sample Name</th>
<th>Group</th>
<th>Sample Date</th>
<th>Sample Time</th>
<th>Sample Elev.</th>
<th>Median Elev.</th>
<th>Percent Glacier</th>
<th>δ¹⁸O M.S.</th>
<th>δ¹⁸O LGR</th>
<th>δ²H LGR</th>
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<td>17</td>
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<td>3915</td>
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<td>-15.55</td>
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<td>27</td>
<td>Dem conf</td>
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<td>14</td>
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<tr>
<td>28</td>
<td>Anc conf</td>
<td>Main Channel</td>
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<td>3918</td>
<td>4932</td>
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<td>-16.14</td>
<td>-16.06</td>
<td>-116.4</td>
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<tr>
<td>29</td>
<td>Llan lakes in</td>
<td>Main Channel</td>
<td>07172011</td>
<td>14</td>
<td>3880</td>
<td>4833</td>
<td>33</td>
<td>-16.17</td>
<td>-15.92</td>
<td>-115.5</td>
</tr>
<tr>
<td>30</td>
<td>Llan lakes out</td>
<td>Main Channel</td>
<td>07172011</td>
<td>14</td>
<td>3850</td>
<td>4793</td>
<td>29</td>
<td>-15.88</td>
<td>-15.85</td>
<td>-115.6</td>
</tr>
</tbody>
</table>
Table 4.2: Synoptic sample data from Mark and McKenzie (2007). Watershed locations are show in Figure 4.1.

<table>
<thead>
<tr>
<th>Sample Name</th>
<th>Sample Elevation m a.s.l.</th>
<th>Median Watershed Elevation m a.s.l.</th>
<th>Watershed Area km²</th>
<th>Percent Glacier Cover 2004</th>
<th>δ¹⁸O 2004</th>
<th>δ¹⁸O 2005</th>
<th>δ²H 2004</th>
<th>δ²H 2005</th>
</tr>
</thead>
<tbody>
<tr>
<td>Querococha (Q3)</td>
<td>3980</td>
<td>4523</td>
<td>64.7</td>
<td>3.4</td>
<td>-13.51</td>
<td>-13.38</td>
<td>-102.6</td>
<td>-100.3</td>
</tr>
<tr>
<td>Pachacoto</td>
<td>3765</td>
<td>4653</td>
<td>202.2</td>
<td>6.3</td>
<td>-13.56</td>
<td>-14.04</td>
<td>-106.3</td>
<td>-103.9</td>
</tr>
<tr>
<td>Yanayacu</td>
<td>3705</td>
<td>4419</td>
<td>268.0</td>
<td>3.7</td>
<td>-13.38</td>
<td>-13.78</td>
<td>-102.5</td>
<td>-102.6</td>
</tr>
<tr>
<td>Olleros</td>
<td>3540</td>
<td>4428</td>
<td>171.6</td>
<td>9.2</td>
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<td>-14.34</td>
<td>-99.2</td>
<td>-104.7</td>
</tr>
<tr>
<td>Quilcay</td>
<td>3150</td>
<td>4589</td>
<td>245.1</td>
<td>16.0</td>
<td>-14.06</td>
<td>-14.61</td>
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<td>-107.8</td>
</tr>
<tr>
<td>Pariac</td>
<td>3144</td>
<td>4207</td>
<td>106.6</td>
<td>10.7</td>
<td>-14.25</td>
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<td>Marcara</td>
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<td>Paltay (Ischinca)</td>
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<td>4545</td>
<td>85.5</td>
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<td>-13.79</td>
<td>-14.32</td>
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<td>-103.5</td>
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<tr>
<td>Buin</td>
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<td>4330</td>
<td>167.3</td>
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<td>-105.8</td>
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<tr>
<td>Llullan</td>
<td>2350</td>
<td>4487</td>
<td>142.2</td>
<td>19.3</td>
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<td>-103.4</td>
</tr>
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<td>4161</td>
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<tr>
<td>Rio Santa 1</td>
<td>3800</td>
<td>4206</td>
<td>288.4</td>
<td>0.9</td>
<td>-12.26</td>
<td>-12.23</td>
<td>-99.3</td>
<td>-95.0</td>
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<td>Rio Santa 2</td>
<td>3462</td>
<td>4256</td>
<td>1635.3</td>
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<td>-13.59</td>
<td>-103.4</td>
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<tr>
<td>Rio Santa Jangas</td>
<td>2807</td>
<td>4221</td>
<td>2567.5</td>
<td>5.0</td>
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<td>-14.02</td>
<td>-100.2</td>
<td>-101.7</td>
</tr>
<tr>
<td>Rio Santa Low</td>
<td>2000</td>
<td>4117</td>
<td>4492.8</td>
<td>7.0</td>
<td>-13.50</td>
<td>-101.8</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
Table 4.3: Results of the two-component isotopic mixing model using meltwater and groundwater sample subsets. Melt and groundwater averages are shown along with one standard deviation from the mean.

<table>
<thead>
<tr>
<th>Main Channel Sites</th>
<th>Main Channel Average (n=2)</th>
<th>Glacier Melt Average (n=5)</th>
<th>Groundwater Average (n = 10)</th>
<th>Percent Glacier Melt</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>$\delta^{18}$O</td>
<td>$\delta^2$H</td>
<td>$\delta^{18}$O</td>
<td>$\delta^2$H</td>
</tr>
<tr>
<td>Llan Lakes In &amp; Llan Lakes Out</td>
<td>-15.89±0.04</td>
<td>-115.6</td>
<td>-16.28</td>
<td>-117.8</td>
</tr>
</tbody>
</table>
4.8 Figures

**Figure 4.1:** The Rio Santa watershed and locations of synoptic sampling points from Mark and McKenzie (2007), as well as associated watersheds. Glacier outlines derived from an August 2010 Landsat TM scene are also shown.
Figure 4.2: Map of the Llanganuco watershed with sample locations shown. Sample numbers correspond to sample names in Table 1.
Figure 4.3: Comparison of δ¹⁸O values measured by a mass spectrometer and Los Gatos instrument.
Figure 4.4: The relationship between $\delta^{18}$O and $\delta^2$H for Llanganuco samples from this study as well as samples collected by Bryan Mark (pers. comm.) between 2004 and 2011.

Figure 4.5: The relationship between $\delta^{18}$O and $\delta^2$H for synoptic samples collected by Bryan Mark between July 2004 and July 2007. Local meteoric water lines are fit through each sample group.
Figure 4.6: Sample median watershed elevation versus $\delta^{18}$O (a) and versus $\delta^2$H (b). Trend lines are only fit through data points that are part of the groups Main Channel and Melt Tribs.

Figure 4.7: Percent glacier cover versus $\delta^{18}$O (a) and versus $\delta^2$H (b) for each Llanganuco sample. Trend lines are only fit through data points that are part of the groups Main Channel and Melt Tribs.
Figure 4.8: Median watershed elevation versus percent glacier cover for each sample. The trend line is fit through all data points.
Figure 4.9: Median watershed elevation plotted versus $\delta^{18}$O (a) and versus $\delta^{2}H$ (b) for each synoptic sample from the 2004 and 2005 dry season (excluding 0% glacierized Conococha).

Figure 4.10: Percent glacier cover for each synoptic watershed plotted versus $\delta^{18}$O (a) and versus $\delta^{2}H$ (b) for each synoptic sample from the 2004 and 2005 dry season (excluding 0% glacierized Conococha).
Figure 4.11: Monthly precipitation and isotopic values of samples collected in the dry season from 2004 to 2011(a). Annual precipitation preceding the date of sampling and the isotopic value for that year are also shown to illustrate the rainout effect for $\delta^{18}$O (b) and $\delta^2$H (c).

Figure 4.12: Boxplots showing variation of each sample group for $\delta^{18}$O (a) and $\delta^2$H (b).
Figure 4.13.a: Average monthly precipitation totals and average monthly $\delta^{18}$O values from 2006 to 2008 for the Marcapomacocha station, Peru (11.4°S, 76.3°W, 4477 m a.s.l.).

b. Average monthly precipitation totals and average monthly $\delta^{18}$O values from 1996 to 2006 for the La Paz station, Bolivia (16.5°S, 68.1°W, 3635 m a.s.l.). Error bars for each data series represent one standard deviation.
5 Conclusions
The majority of glaciers around the world are losing area and mass. While the physical reasons for these changes vary from one region to another, the lasting societal effects will be similar. Glacier mass loss will continue to contribute to global sea level rise. Furthermore, populations which are adjacent to these natural water towers will experience declines in water availability as glacier storage decreases. In this thesis I focused on a region of the world that is particularly vulnerable, the Cordillera Blanca, Peru. Over a quarter of a million people live in the regional Callejón de Huaylas watershed that is most adjacent to the this mountain range. Hydroelectric power plants, agricultural water diversions, and the large coastal city of Chimbote all sit further downstream, awaiting the water that originates in the high Andes. Most of this thesis focused on a single, highly glacierized watershed of the Cordillera Blanca, Llanganuco, which is taken to be representative of the Cordillera Blanca.

In response to the initial research questions outlined in the Introduction, we discuss the associated conclusions from this thesis.

1. Glacier area extent decreased significantly in the Cordillera Blanca from 1987 to 2010. Between these years glacier area in the entire Cordillera Blanca decreased by 25%, while glacier area in the Llanganuco watershed decreased by 19%. The rate of glacier area loss is accelerating based on our most recent estimates of glacier area change.
2. The application of image atmospheric correction is important when measuring glacier area with satellite images. We observed a 5% difference in the debris-free glacier area calculated with an atmospherically-corrected and non-atmospherically corrected Landsat scene. Atmospheric correction does not appear to affect relative changes in glacier area. However, it is more important for determining accurate glacier areas at a single point in time.
3. The choice of threshold associated with automated glacier mapping methods is very important. In order to eliminate error in change analyses, analysts should
atmospherically correct each image and then use a single, validated threshold value.

4. The updated DHSVM, with a newly-coupled dynamic glacier model, performs relatively well based on calibration and validation with monthly discharge and sub-decadal estimates of glacier area. Our model efficiency criteria are acceptable, but less than ideal for monthly calibration. It was not possible for us to quantify our total uncertainty due to the large amount of uncertainty associated with the many individual input parameters. Furthermore, we lack rigorous on-the-ground validation data which is necessary when using a complex model such as DHSVM. Therefore, the model’s predictive validity is lacking and it cannot yet be used to make predictions about future discharge associated with different climate scenarios.

5. We collected stable isotope samples to better understand some of the hydrologic processes occurring in the Llanganuco watershed. In this watershed, isotopic variability is best explained by percent glacier cover of the subwatershed associated with the water sample. This trend is persistent when scaled up to the larger watersheds of Cordillera Blanca. While these findings are still observational in nature, they could be used to more accurately determine changing contributions of glacier meltwater on a regional scale, provided that temporal variations of other source waters are known.

6. We used a simple two-component isotopic mixing model to estimate relative contributions of glacier meltwater and groundwater during a point in time of the 2011 dry season. We estimate that approximately three-fourths of surface water leaving the watershed is derived from glacier meltwater, but our estimates have large error margins. Further work is needed to refine these estimates.

Although we have answered a number of research questions and shed light on some aspects of glacier change and hydrology in the Cordillera Blanca, there are still many questions that remain. In terms of glacier mapping, it will be important to determine a robust method for estimating the error associated with a glacier outline. In this study we
used the one pixel buffer method which tends to overestimate error. Another current challenge is associated with mapping debris-covered glaciers using satellite imagery. Automated methods need to be developed so that these glaciers may be mapped accurately and efficiently. More work also needs to be done in order to understand the hydrologic role of debris-covered glaciers. We neglected debris-covered glaciers in our model, but acknowledge that doing so may have been invalid. In the Cordillera Blanca there is simply not enough data related to the behavior of these glaciers.

Even though statistical and hybrid-type hydrological models have been utilized successfully in the Cordillera Blanca, we argue that distributed, physical models which include glacier dynamics are the way forward. Since models like the DHSVM include glacier dynamics, it is possible to more accurately predict temporal changes in glacier size and the resulting impacts on discharge. But currently, these models are still very difficult to apply due to data availability limitations. Therefore, we advocate for continued installation of micro-meteorological stations and additional field measurements of glacier mass balance. Future satellite missions may help to alleviate some of these data limitations.

The stable isotopes of water appear to be promising tools for helping to understand the contribution of glacier meltwater to streamflow on sub-watershed to regional scales. Before we can accurately estimate this contribution on an annual and seasonal basis we need to understand the spatial and temporal variability of different source waters. This means that glacier meltwater, groundwater, and precipitation need to be sampled throughout the year and across watersheds with differing percent glacier cover. Along the same lines, it will be very important to quantify different sources of groundwater recharge as well as groundwater resonance times in watersheds of the Cordillera Blanca.
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APPENDICES
Appendix A: Step-by-step glacier mapping methods


2. Open the satellite files in ENVI and visually inspect the scene and histogram.

3. Atmospherically correct with 6S if the scene is of good quality

4. Stack bands in ENVI and create a false color image.

5. Perform desired glacier mapping band ratios and save as geotiff files.
   a. \( B3 \text{div} B5 = \frac{\text{band3}}{\text{band5}} \) for Landsat TM
   b. \( \text{NDGI} = \frac{(\text{band4}-\text{band5})}{(\text{band4}+\text{band5})} \) for Landsat TM
   c. \( \text{NDSI} = \frac{(\text{band2}-\text{band5})}{(\text{band2}+\text{band5})} \) for Landsat TM

6. Start working in ArcGIS because Arc is better for dealing with vector data. Load the false color raster (usually B543 combination) and the band ratio raster.

7. Using the SRTM DEM and the false color raster, find the approximate minimum terminus elevation for the earliest year you’ll map. Subtract a few hundred meters from this value.
   a. Create an elevation mask for the watershed or area of interest by using the raster calculator. (ex: \( \text{DEM} >= 2000 \text{m} \))
   b. Convert the resulting raster calculation to a polygon (Conversion tools >> From Raster >> to Polygon). This will be used to clip the band ratio later.

8. Determine a threshold for your particular band ratio. Pixels below the threshold are non-ice (data value = NoData) and pixels above the threshold are mapped as ice (data value = 1).

9. Using Raster calculator, select pixels greater than or equal to the determined threshold for the band ratio raster and pixels that are greater than the threshold elevation for the DEM. For example: \( b3 \text{divb5} >= 3 \) & \( \text{DEM} >= 2000 \)
137

a. Make this calculation permanent with the file name thr#_gt####m.tif. For example the b3divb5 threshold was 3 and the elevation threshold was 2000m so we would save the file as thr3_gt2000m.tif

10. Next, reclassify thr#_gt####m.tif so that it only has values of NoData and 1. Values of 0 should be changed to NoData and values of 1 should remain 1. Save the reclassification as thr#_gt####m_rec.tif

11. Convert the reclassified raster to a polygon with an area feature. To do this, first create a Personal File Geodatabase (call it areas_yyyy.gdb) under mmddyyyy >> Processing >> Glacier_Mapping >> b3divb5. Use the conversion tool (Conversion tools >> From Raster >> to Polygon). Do not select “Simplify Polygons” and be sure to save this shapefile in the new Personal File Geodatabase (areas_yyyy.gdb). Doing this automatically creates an area field for each polygon.

12. Open the attribute table for the newly created set of glacier polygons. Select by Attributes so that only polygons with areas greater than 10000 m² are selected. This step should eliminate most of the misclassified snow bodies. Close the attribute table and right click on the file. Go to Data >> Export Data … Save the selected features using the following format thr#_poly_area_gt10000m2

13. Thus far we have eliminated lakes and shadows below two thresholds. However there may still be lakes and shadows above the elevation threshold. I created a shapefile for all of the lakes I could find in the study area and merged this shapefile with the basin extent file using the Union command. From the merged polygon file, I selected the largest polygon which is the basin extent polygon with lake holes cut out. I called this “Bow_entire_mask_nolakes.shp”. I then used the Clip tool to clip out lakes from thr#_poly_area_gt10000m2. I saved the lake-free file as thr3_poly_area_gt10000m2_NL.shp (where NL signifies no lakes).

14. There are still issues with shadows and new lakes forming in subsequent years. The best way to handle these errors is by visually inspecting the entire extent (granted the area is not huge) for obvious errors. BE CONSISTENT between years.
For more information about glacier mapping I would recommend the GLIMS tutorial page: http://www.glims.org/MapsAndDocs/guides.html as well as the following book:

Appendix B: Values and sources for various DHSVM parameters found in the configuration file

Appendix B.1: Values used in the Constants section of the configuration file

########################################################################
# CONSTANTS SECTION
########################################################################
[CONSTANTS]                               # Model constants
Ground Roughness     = 0.01               # Roughness of soil surface (m)
Snow Roughness       = 0.005               # Roughness of snow surface (m)
Rain Threshold       = 0.5               # Minimum temperature at which rain
          # occurs (C)
Snow Threshold       = 3.3                # Maximum temperature at which snow
          # occurs (C)
Snow Water Capacity  = 0.05               # Snow liquid water holding capacity
          # (fraction)
Reference Height     = 20.0               # Reference height (m)
Rain LAI Multiplier  = 0.0001             # LAI Multiplier for rain interception
Snow LAI Multiplier  = 0.0005             # LAI Multiplier for snow interception
Min Intercepted Snow = 0.005              # Intercepted snow that can only be
          # melted (m)
Outside Basin Value  = 0                  # Value in mask that indicates outside
          # the basin
Temperature Lapse Rate = -0.0065        # Temperature lapse rate (C/m)
Precipitation Lapse Rate = 0.0005        # Precipitation lapse rate (m/m)

Sources for different constant parameter values

Ground Roughness: Brutsaert (2005), Table 2.6, pg 45

Snow Roughness: DeWalle and Rango (2008), p. 161

Rain/Snow Threshold: Dai (2008)

Snow Water Capacity: Singh and Singh (2001), Table 3.4, pg 108

Reference Height: should be 10 m higher than highest vegetation

Rain LAI Multiplier: Mentioned in Andreadis et al. (2009); also Safeeq and Fares (2011); Brutsaert (2005), pg 104
Snow LAI Multiplier: Mentioned in Andreadis et al. (2009)

Min Intercepted Snow: Snow interception not expected to be important

Temperature Lapse Rate: Standard temperature lapse rate

Precipitation Lapse Rate: Roughly based on estimates from Juen et al. (2007)

Note: DHSVM PLR is calculated as

$LappedPrecip = Precip * (1.0 + PrecipLapse * (ToElev - FromElev))$
Appendix B.2: Land cover (vegetation) classes and values used in this study

######################################## VEGETATION 1########################################
Vegetation Description 1 = Woodland (Polylepis)
Overstory Present 1 = TRUE
Understory Present 1 = TRUE
Fractional Coverage 1 = 0.75
Trunk Space 1 = 0.2
Aerodynamic Attenuation 1 = 0.5
Radiation Attenuation 1 = 0.5
Hemi Fract Coverage 1 = 0.8
Clumping Factor 1 =
Leaf Angle A 1 =
Leaf Angle B 1 =
Scattering Parameter 1 =
Max Snow Int Capacity 1 = 0.003
Mass Release Drip Ratio 1 = 0.4
Snow Interception Eff 1 = 0.6
Impervious Fraction 1 = 0.0
Height 1 = 20.0 0.5
Maximum Resistance 1 = 5000. 600.
Moisture Threshold 1 = 0.33 0.13
Vapor Pressure Deficit 1 = 4000 4000
Rpc 1 = .108 .108
Number of Root Zones 1 = 3
Root Zone Depths 1 = 0.10 0.20 0.10
Overstory Root Fraction 1 = 0.20 0.40 0.40
Understory Root Fraction 1 = 0.40 0.60 0.00
Overstory Monthly LAI 1 = 5.0 5.0 5.0 5.0 5.0 4.5 4.0 4.5 5.0 5.0 5.0
Understory Monthly LAI 1 = 1.7 1.7 1.7 1.7 1.7 1.7 1.7 1.7 1.7 1.7 1.7
Overstory Monthly Alb 1 = 0.18 0.18 0.18 0.18 0.18 0.18 0.18 0.18 0.18 0.18 0.18
Understory Monthly Alb 1 = 0.15 0.15 0.15 0.15 0.15 0.15 0.15 0.15 0.15 0.15 0.15

######################################## VEGETATION 2########################################
Vegetation Description 2 = Pampa
Overstory Present 2 = FALSE
Understory Present 2 = TRUE
Fractional Coverage 2 =
Trunk Space 2 =
Aerodynamic Attenuation 2 =
Radiation Attenuation 2 =
Hemi Fract Coverage  2 =
Clumping Factor    2 =
Leaf Angle A       2 =
Leaf Angle B       2 =
Scattering Parameter  2 =
Max Snow Int Capacity  2 =
Mass Release Drip Ratio  2 =
Snow Interception Eff  2 =
Impervious Fraction  2 = 0.0
Height             2 = 0.25
Maximum Resistance  2 = 600
Minimum Resistance  2 = 200
Moisture Threshold  2 = 0.33
Vapor Pressure Deficit  2 = 4000
Rpc                2 = .108
Number of Root Zones  2 = 3
Root Zone Depths    2 = 0.10 0.15 0.2
Overstory Root Fraction  2 =
Understory Root Fraction  2 = 0.60 0.40 0.00
Overstory Monthly LAI  2 =
Understory Monthly LAI  2 = 1.8 1.8 1.8 1.8 1.8 1.8 1.8 1.8 1.8 1.8
Overstory Monthly Alb  2 =
Understory Monthly Alb  2 = 0.15 0.15 0.15 0.15 0.15 0.15 0.15 0.15 0.15 0.15
0.15

############################ VEGETATION 3########################################
Vegetation Description   3 = Water
Overstory Present        3 = FALSE
Understory Present       3 = FALSE
Fractional Coverage      3 =
Trunk Space              3 =
Aerodynamic Attenuation  3 =
Radiation Attenuation    3 =
Hemi Fract Coverage      3 =
Clumping Factor          3 =
Leaf Angle A             3 =
Leaf Angle B             3 =
Scattering Parameter     3 =
Max Snow Int Capacity    3 =
Mass Release Drip Ratio  3 =
Snow Interception Eff    3 =
Impervious Fraction      3 = 0.0
Height                   3 =
Maximum Resistance       3 =
Minimum Resistance 3 =
Moisture Threshold 3 =
Vapor Pressure Deficit 3 =
Rpc 3 =
Number of Root Zones 3 = 3
Root Zone Depths 3 = 0.10 0.10 0.10
Overstory Root Fraction 3 =
Understory Root Fraction 3 = 0.00 0.00 0.00
Overstory Monthly LAI 3 =
Understory Monthly LAI 3 =
Overstory Monthly Alb 3 =
Understory Monthly Alb 3 =

VEGETATION 4
Vegetation Description 4 = Glacier
Overstory Present 4 = FALSE
Understory Present 4 = FALSE
Fractional Coverage 4 =
Trunk Space 4 =
Aerodynamic Attenuation 4 =
Radiation Attenuation 4 =
Hemi Fract Coverage 4 =
Clumping Factor 4 =
Leaf Angle A 4 =
Leaf Angle B 4 =
Scattering Parameter 4 =
Max Snow Int Capacity 4 =
Mass Release Drip Ratio 4 =
Snow Interception Eff 4 =
Impervious Fraction 4 = 0.0
Height 4 =
Maximum Resistance 4 =
Minimum Resistance 4 =
Moisture Threshold 4 =
Vapor Pressure Deficit 4 =
Rpc 4 =
Number of Root Zones 4 = 3
Root Zone Depths 4 = 0.10 0.10 0.10
Overstory Root Fraction 4 =
Understory Root Fraction 4 = 0.0 0.0 0.0
Overstory Monthly LAI 4 =
Understory Monthly LAI 4 =
Overstory Monthly Alb 4 =
Understory Monthly Alb 4 = 0.35 0.35 0.35 0.35 0.35 0.35 0.35 0.35 0.35 0.35
0.35

############################ VEGETATION 5############################################
Vegetation Description 5 = Bare
Overstory Present 5 = FALSE
Understory Present 5 = FALSE
Fractional Coverage 5 =
Trunk Space 5 =
Aerodynamic Attenuation 5 =
Radiation Attenuation 5 =
Hemi Fract Coverage 5 =
Clumping Factor 5 =
Leaf Angle A 5 =
Leaf Angle B 5 =
Scattering Parameter 5 =
Max Snow Int Capacity 5 =
Mass Release Drip Ratio 5 =
Snow Interception Eff 5 =
Impervious Fraction 5 = 0.0
Height 5 =
Maximum Resistance 5 =
Minimum Resistance 5 =
Moisture Threshold 5 =
Vapor Pressure Deficit 5 =
Rpc 5 =
Number of Root Zones 5 = 3
Root Zone Depths 5 = 0.10 0.15 0.2
Overstory Root Fraction 5 =
Understory Root Fraction 5 = 0.0 0.0 0.00
Overstory Monthly LAI 5 = 0.0 0.0 0.0 0.0 0.0 0.0 0.0 0.0 0.0 0.0 0.0 0.0
Understory Monthly LAI 5 = 0.0 0.0 0.0 0.0 0.0 0.0 0.0 0.0 0.0 0.0 0.0 0.0
Overstory Monthly Alb 5 = 0.00 0.00 0.00 0.00 0.00 0.00 0.00 0.00 0.00 0.00 0.00 0.00
Understory Monthly Alb 5 = 0.00 0.00 0.00 0.00 0.00 0.00 0.00 0.00 0.00 0.00 0.00 0.00
0.00

############################ VEGETATION 6############################################
Vegetation Description 6 = Grassland
Overstory Present 6 = FALSE
Understory Present 6 = TRUE
Fractional Coverage 6 =
Trunk Space 6 =
Aerodynamic Attenuation 6 =
Radiation Attenuation 6 =
Hemi Fract Coverage 6 =
Clumping Factor 6 =
Leaf Angle A 6 =
Leaf Angle B 6 =
Scattering Parameter 6 =
Max Snow Int Capacity 6 =
Mass Release Drip Ratio 6 =
Snow Interception Eff 6 =
Impervious Fraction 6 = 0.0
Height 6 = 1
Maximum Resistance 6 = 600
Minimum Resistance 6 = 200
Moisture Threshold 6 = 0.33
Vapor Pressure Deficit 6 = 4000
Rpc 6 = .108
Number of Root Zones 6 = 3
Root Zone Depths 6 = 0.10 0.15 0.20
Overstory Root Fraction 6 =
Understory Root Fraction 6 = 0.40 0.60 0.00
Overstory Monthly LAI 6 =
Understory Monthly LAI 6 = 1.8 1.8 1.8 1.8 1.8 1.8 1.8 1.8 1.8 1.8 1.8 1.8
Overstory Monthly Alb 6 =
Understory Monthly Alb 6 = 0.15 0.15 0.15 0.15 0.15 0.15 0.15 0.15 0.15 0.15 0.15 0.15 0.15

################################################################ VEGETATION 7################################################################
Vegetation Description 7 = Shrubland # (Closed)
Overstory Present 7 = FALSE
Understory Present 7 = TRUE
Fractional Coverage 7 =
Trunk Space 7 =
Aerodynamic Attenuation 7 =
Radiation Attenuation 7 =
Hemi Fract Coverage 7 =
Clumping Factor 7 =
Leaf Angle A 7 =
Leaf Angle B 7 =
Scattering Parameter 7 =
Max Snow Int Capacity 7 =
Mass Release Drip Ratio 7 =
Snow Interception Eff 7 =
Impervious Fraction 7 = 0.0
Height 7 = 1.5
Maximum Resistance 7 = 600
Minimum Resistance $= 200$
Moisture Threshold $= 0.33$
Vapor Pressure Deficit $= 4000$
Rpc $= 0.108$
Number of Root Zones $= 3$
Root Zone Depths $= 0.10 \ 0.15 \ 0.20$
Overstory Root Fraction $= 0.40 \ 0.60 \ 0.00$
Overstory Monthly LAI $= 2.0 \ 2.0 \ 2.0 \ 2.0 \ 2.0 \ 2.0 \ 2.0 \ 2.0$
Overstory Monthly Alb $= 0.15 \ 0.15 \ 0.15 \ 0.15 \ 0.15 \ 0.15 \ 0.15 \ 0.15 \ 0.15 \ 0.15 \ 0.15 \ 0.15 \ 0.15 \ 0.15 \ 0.15 \ 0.15$

Sources for different vegetation parameter values

Fractional coverage: July 2011 field estimates

Trunk space: July 2011 field estimates

Aerodynamic Attenuation: Appendix of Wigmosta et. al. (1994)

Radiation Attenuation: July 2011 field estimates, Wigmosta et. al. (1994)

Note: Based on Pascal Storck’s field work, short wave radiation beneath the canopy, $\tau$, can be estimated as 10-20% of that with no canopy

Hemi Fract Coverage: July 2011 field estimates

Clumping Factor: not required

Leaf Angle A: not required

Leaf Angle B: not required

Scattering Parameter: not required

Max Snow Int Capacity: Storck (2000)


Snow Interception Eff: Storck (2000)

Impervious Fraction: field observations

Note: Setting this value to 0 leads the model to require a surface routing file

Height: July 2011 field estimates

Maximum Resistance: Table 1 of Wigmosta et al. (1994)
Minimum Resistance: Table 1 of Wigmosta et al. (1994)

Moisture Threshold: eqn. 16 in Wigmosta et al. (2004); Handbook of Hydrology (1996)

Vapor Pressure Deficit: Wigmosta et. al. (1994)

Rpc: Table 1 of Wigmosta et al. (1994)

Note: Wigmosta et al. (1994) states this is the light level where \( rs = 2rs_{min} \) (equation 12).

Number of Root Zones: 3 is standard for DHSVM

Root Zone Depths: estimated

Overstory Root Fraction: estimated

Understory Root Fraction: estimated

Overstory Monthly LAI: Brutsaert (2005) (Table 2.9, pg 69); Table 1

Note: This can also be based on satellite data. One source for this is the LDAS project (http://ldas.gsfc.nasa.gov/) which is based on the global Myneni (1997) data set.

Understory Monthly LAI: Brutsaert (2005) (Table 2.9, pg 69); Table 1 in Wigmosta et. al. (1994)

Note: This can also be based on satellite data. One source for this is the LDAS project (http://ldas.gsfc.nasa.gov/) which is based on the global Myneni (1997) data set

Overstory Monthly Alb: Dingman (2002), Table D-2, pg 584; Brutsaert (2005), Table 2.7, pg 64; Hydrology Handbook (1996), Table 4.4, pg 133

Understory Monthly Alb: Dingman (2002), Table D-2, pg 584; Brutsaert (2005), Table 2.7, pg 64; Hydrology Handbook (1996), Table 4.4, pg 133
Appendix B.3: Soil classes and values used in this study

############################ SOIL 1#############################
Soil Description 1 = SANDY LOAM
Lateral Conductivity 1 = 0.000347
Exponential Decrease 1 = 2
Maximum Infiltration 1 = 3e-3
Surface Albedo 1 = 0.1
Number of Soil Layers 1 = 3
Porosity 1 = .44 .44 .44
Pore Size Distribution 1 = .20 .20 .20
Bubbling Pressure 1 = .22 .22 .22
Field Capacity 1 = .2 .2 .2
Wilting Point 1 = .07 .07 .07
Bulk Density 1 = 1550. 1550. 1550.
Vertical Conductivity 1 = 0.000347 0.000347 0.000347
Thermal Conductivity 1 = 7.114 6.923 7.0
Thermal Capacity 1 = 1.4e6 1.4e6 1.4e6

############################ SOIL 2#############################
Soil Description 2 = Pampa
Lateral Conductivity 2 = 0.000347
Exponential Decrease 2 = 1
Maximum Infiltration 2 = 3e-3
Surface Albedo 2 = 0.08
Number of Soil Layers 2 = 3
Porosity 2 = .7 .7 .44
Pore Size Distribution 2 = .20 .20 .20
Bubbling Pressure 2 = .22 .22 .22
Field Capacity 2 = .4 .4 .2
Wilting Point 2 = .07 .07 .07
Bulk Density 2 = 1550. 1550. 1550.
Vertical Conductivity 2 = 0.000347 0.000347 0.000347
Thermal Conductivity 2 = 7.114 6.923 7.0
Thermal Capacity 2 = 1.4e6 1.4e6 1.4e6

############################ SOIL 3#############################
Soil Description 3 = WATER (as clay)
Lateral Conductivity 3 = 0.0000000128
Exponential Decrease 3 = 0
Maximum Infiltration 3 = 1e-5
Surface Albedo 3 = 0.08
Number of Soil Layers 3 = 3
Porosity 3 = .48 .48 .48
Pore Size Distribution 3 = .09 .09 .09
Bubbling Pressure 3 = .41 .41 .41
Field Capacity 3 = .34 .34 .34
Wilting Point 3 = .26 .26 .26
Bulk Density 3 = 1300. 1300. 1300.
Vertical Conductivity 3 = 0.0000000128 0.0000000128 0.0000000128
Thermal Conductivity 3 = 7.114 6.923 7.0
Thermal Capacity 3 = 1.4e6 1.4e6 1.4e6

############################ SOIL 4#################################################
Soil Description 4 = Gravels/Bedrock
Lateral Conductivity 4 = 0.00007
Exponential Decrease 4 = 0
Maximum Infiltration 4 = 3e-3
Surface Albedo 4 = 0.55
Number of Soil Layers 4 = 3
Porosity 4 = .4 .4 .4
Pore Size Distribution 4 = .20 .20 .20
Bubbling Pressure 4 = .22 .22 .22
Field Capacity 4 = .2 .2 .2
Wilting Point 4 = .07 .07 .07
Bulk Density 4 = 1550. 1550. 1550.
Vertical Conductivity 4 = 0.00007 0.00007 0.00007
Thermal Conductivity 4 = 7.114 6.923 7.0
Thermal Capacity 4 = 1.4e6 1.4e6 1.4e6

############################ SOIL 5#################################################
Soil Description 5 = LOAMY SAND
Lateral Conductivity 5 = 0.000156
Exponential Decrease 5 = 2
Maximum Infiltration 5 = 6.0e-3
Surface Albedo 5 = 0.15
Number of Soil Layers 5 = 3
Porosity 5 = .41 .41 .41
Pore Size Distribution 5 = .23 .23 .23
Bubbling Pressure 5 = .09 .09 .09
Field Capacity 5 = .2 .2 .2
Wilting Point 5 = .06 .06 .06
Bulk Density 5 = 1600. 1600. 1600.
Vertical Conductivity 5 = 0.000156 0.000156 0.000156
Thermal Conductivity 5 = 7.114 6.923 7.0
Thermal Capacity 5 = 1.4e6 1.4e6 1.4e6
Sources for different soil parameter values.

Soil Description: based on July 2011 field estimates

Lateral Conductivity: Dingman (2002), Table 6-1, pg 235 which is originally from Clapp and Hornberger (1978);
www.mo10.nrcs.usda.gov/references/guides/properties/lithicperms.html;
also see Schwartz and Zang (2003)

Exponential Decrease: Beven (1982); mentioned in Wigmosta et al. (1994); Niu et al. (2005); Yao and Yang (2009); Also see Whitaker et al. (2003) and Wang, et al. (2006)

Maximum Infiltration: http://www.fao.org/docrep/S8684E/s8684e0a.htm

Surface Albedo: Brutsaert (2005), Table 2.7, pg 64

Number of Soil Layers: estimated

Porosity: Dingman (2002), Table 6-1, pg 235 which is originally from Clapp and Hornberger (1978)

www.mo10.nrcs.usda.gov/references/guides/properties/lithicperms.html;
also see Schwartz and Zang (2003)

Pore Size Distribution: Dingman (2002), Table 6-1, pg 235 which is originally from Clapp and Hornberger (1978)

Bubbling Pressure: Dingman (2002), Table 6-1, pg 235 which is originally from Clapp and Hornberger (1978)

Field Capacity: Meyer et al. (1997), Appendix A.4; Also see Dingman (2002), pg 225

Wilting Point: Meyer et al. (1997), Appendix A.4; Also see Dingman (2002), pg 225

Bulk Density:
/www.pedosphere.com/resources/bulkdensity/triangle_us.cfm?49,308

Vertical Conductivity: Dingman (2002), Table 6-1, pg 235 which is originally from Clapp and Hornberger (1978);
www.mo10.nrcs.usda.gov/references/guides/properties/lithicperms.html;
also see Schwartz and Zang (2003)
Thermal Conductivity: Wigmosta et al. (1994)
Thermal Capacity: Wigmosta et al. (1994)
Appendix B.4: Bibliography for Appendix B


Storck, P., 2000, Trees, snow and flooding: an investigation of forest canopy effects on snow accumulation and melt at the plot and watershed scales in the Pacific Northwest: Washington State University Department of Civil and Environmental Engineering


Appendix C: Downloading and Bias Correcting Raw MERRA data

Appendix C.1: Downloading MERRA data and creating a forcing file

The MERRA data holdings page (http://disc.sci.gsfc.nasa.gov/daac-bin/DataHoldings.pl) contains information on different methods for downloading data. For all but two variables, we used the Data Subsetter option. We needed to use the Simple Subset Wizard to download the wind speed variable because this variable was not available with the Data Subsetter option. We used the Mirador option to download the surface geopotential field which was used for calculating grid cell elevation. For each variable, except surface geopotential, we downloaded 1 hour-averaged data for two dimensions only. The beginning of the Standard Name Legend associated with all of the products, except the surface geopotential field, is “tavg1_2d”. The next portion of the Standard Name Legend is the group that the specific variable falls into. We downloaded variables from the following groups: surface turbulent fluxes and related quantities (“flx”), radiation (“rad”), and single level (“slv”). The final part of the Standard Name Legend is related to the horizontal and vertical resolution. All products were downloaded at native (2/3 x 1/2 deg) horizontal resolution (“N”) and two-dimensional (“x”) vertical resolution. Below we list the different data products that were downloaded and the specific fields that were downloaded.

- tavg1_2d_flx_Nx
  - PRECTOT = Total surface precipitation flux (kg m⁻²)
    - Downloaded with Data Subsetter
  - SPEED = Effective surface wind speed including 3d winds and gustiness (m s⁻¹)
    - Downloaded with Simple Subset Wizard

- tavg1_2d_rad_Nx
  - LWGAB = Absorbed longwave at the surface (W m⁻²)
    - Downloaded with Data Subsetter
- SWGDN = Surface incident shortwave flux (W m⁻²)
  - Downloaded with Data Subsetter
- tavg1_2d_slv_Nx
  - T2M = Temperature at 2 m above the displacement height (K)
    - Downloaded with Data Subsetter
  - QV2M = Specific humidity at 2 m above the displacement height (kg kg⁻¹)
    - Downloaded with Data Subsetter
- const_2d_mld_Nx
  - PHIS - surface geopotential height (m² s⁻²)
    - Downloaded using Mirador

We downloaded data for 1980 to 2011. We used an area of interest within the Llanganuco watershed to subset the data (West: -77.646°, North: -9.018°, South: -9.090°, East: -77.570°). Since our area of interest is much smaller than the size of a MERRA grid cell the Data Subsetter option subsets the data to include the four closest grid cells to our area of interest. The Simple Subset Wizard, on the other hand, only subsets the data to include the nearest grid cell point if the area of interest is completely within a grid cell. All files were downloaded in .hdf format using the wget script (http://ftp.gnu.org/gnu/wget/) for UNIX. Time series for each variable were creating using a Matlab script.

The digital elevation model derived from the native resolution surface geopotential height file is shown below.
Figure C.1: MERRA grid cell surface elevation. The yellow (GP 1) and red (GP 2) grid cells are the two pixels closest to the Llanganuco watershed.
Appendix C.2: Bias correction of the raw MERRA data

We compared average monthly values of temperature, wind speed, relative humidity, and incoming shortwave radiation from station observations (Llan A) and MERRA data for the period 2004 to 2010 (Figure C.2).

**Figure C.2:** Average monthly values (2004 to 2010) from the meteorological station Llan A compared with average monthly values from the same period for MERRA grid points 1 and 2 (see Figure C.1).

We also compared the average monthly values of incoming longwave radiation from observations at Artesonraju glacier (Juen, 2006), at an elevation of 4850 m, and MERRA data (Figure C.3).
Figure C.3: Average monthly values of longwave radiation (2004 to 2005) from a meteorological station located in the Paron watershed (4850 m) (Juen, 2006) compared with average monthly values from the same period for MERRA grid points 1 and 2 (see Figure C.1).

Next, we compared average temperature (Figure C.4), wind speed (Figure C.5), relative humidity (Figure C.6), shortwave radiation (Figure C.7), and longwave radiation (Figure C.8) from MERRA with station observations for each hour of each month. Data used for the first four variables came from Llan A (2004 to 2010), while the longwave comparison used data from Artesonraju glacier (April 2004 to March 2005) (Juen, 2006).

We also compared average monthly totals of precipitation from station data and both MERRA grid points, GP1 and GP2 (Figure C.9).
Figure C.4: Average hourly values of temperature for each month measured at station Llan A and extracted from MERRA grid points 1 and 2.
Figure C.5: Average hourly values of wind speed for each month measured at station Llan A (2004 to 2010) and extracted from MERRA grid points 1 and 2.
Figure C.6: Average hourly values of relative humidity for each month measured at station Llan A (2004 to 2010) and extracted from MERRA grid points 1 and 2.
Figure C.7: Average hourly values of shortwave radiation for each month measured at station Llan A (2004 to 2010) and extracted from MERRA grid points 1 and 2.
Figure C.8: Average hourly values of longwave radiation for each month measured at a station in the Paron watershed (2004 to 2005) and extracted from MERRA grid points 1 and 2.
Figure C.9: Average monthly precipitation totals measured at station Llan A (2004 to 2010) and extracted from MERRA grid points 1 and 2.