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John G Van Hoesen
University of Nevada, Las Vegas

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Doctor of Philosophy in Geoscience

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ABSTRACT

Late Quaternary Glacial and Periglacial Environments, Snake Range, Nevada

by

John G. Van Hoesen

Dr. Richard L. Orndorff, Examining Committee Chair
Assistant Professor of Geology
Eastern Washington University

Limited research has been conducted on the paleoclimatic significance of glacial and periglacial features in the Great Basin. Glacial features in the range were first recognized and described by early explorers (Gilbert, 1875; Simpson, 1876 and Russell, 1884) and subsequent authors have continued to substantiate and elaborate on earlier reports (Heald, 1956; Kramer, 1962; Currey, 1969; Peigat, 1980; Osborn and Bevis, 2001). Since this early reconnaissance work, few studies have focused on the Late Quaternary evolution and paleoclimatic implications of glacial and periglacial landforms in the Great Basin (Wayne, 1983; Osborn, 1989; Bevis, 1995; and Osborn and Bevis, 2001). Wayne (1984) describes relict rock glaciers, sorted circles, debris islands, solifluction lobes and sorted stripes in the Ruby, Schell Creek, and Snake ranges. While Currey (1969) and Osborn and Bevis, (2001) discuss the distribution of rock glaciers in numerous ranges throughout the Great Basin, including the Snake Range, and the surrounding regions.
This study presents new data on the glacial and periglacial Late Quaternary conditions in the interior Great Basin based on studies carried out in the Snake Range, located in east-central Nevada. I propose the Lehman rock glacier is an ice-cemented landform that evolved via a recessional genesis, contrary to present glacial or periglacial models that primarily propose constructional geneses for rock glaciers. Preliminary GPR evidence suggests the Lehman rock glacier may retain interstitial lenses of ice; remnant ice that has stagnated under modern climate conditions.

The spatial distribution of both glacial and periglacial landforms provides paleoclimatic information derived using field evidence and computer modeling. Neoglacial temperature depression estimates calculated using modern freezing and thawing indices for relict rock glaciers, range from −0.25 °C to −1.00 °C while temperature estimates calculated using the methodology described by Frauenfelder and Kääb, (2000) and Frauenfelder et al., (2001) are 0.35 to 0.97 degrees lower than those calculated using freezing and thawing indices. Full Glacial MAAT depression estimates range from approximately −5.16°C to −6.61°C calculated using periglacial landforms, to approximately −4.55°C to −5.77°C, calculated from reconstructed Angel Lake equilibrium line altitudes (ELAs).

Finally, using scanning electron microscopy, it is possible to differentiate between a glacial and non-glacial origin for pebbles and cobbles entrained in enigmatic sedimentary deposits is a useful tool. Each depositional environment creates distinct micro-features that can be identified and used to establish a glacial or non-glacial history.
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CHAPTER 1

SIGNIFICANCE OF LATE QUATERNARY LANDFORMS IN THE INTERIOR GREAT BASIN OF THE SOUTHWESTERN UNITED STATES

The Great Basin is located in the Basin and Range physiographic province that drains internally (e.g. water that enters the Great Basin undergoes evaporation or groundwater recharge and never reaches the ocean via fluvial drainages). Centered on the state of Nevada, the Basin and Range Province is an extensive region (~ 200,000 mi^2) of alternating, north-south trending, faulted mountains and flat valley floors (Fig. 1.1). The entire area experienced Cenozoic extension that caused the crust to deform and thin as it was pulled apart, producing in extensive normal faulting. Numerous mountain ranges were uplifted and valleys down-dropped along these roughly north-south-trending faults, creating the characteristic horst and graben topography of the Basin and Range province. The southern Snake Range is one such uplifted horst located in east central Nevada (Fig. 1.1). This range is composed of early Paleozoic quartzite, limestone, and shale intruded by the Jurassic Snake Creek-Williams Canyon Pluton (Drewes, 1958; Whitebread, 1969; Lee et al., 1970; Lee and Van Loenen, 1971; Lee et al., 1981; Lee and Christiansen, 1983 a, b; Lee et al., 1986; McGrew and Miller, 1995 and Miller, 1999). The dominant lithologies in the southern Snake Range are Cambrian aged Prospect Mountain quartzite and Middle Cambrian Pole Canyon limestone and Pioche Shale.

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Figure 1.1: Line diagram illustrating the geographic extent of the Great Basin Physiographic Province in the Western United States.
Topographically high regions like the Snake Range are found throughout the Great Basin and produce drastically different microclimates from the surrounding lower elevation valleys. During the Late Quaternary climatic conditions in the Snake Range, and other similar alpine environments throughout the Great Basin, facilitated the development of many glacial and periglacial environments, which are no longer supported under modern climate.

It has long been recognized that the Snake Range experienced glaciation during the Pleistocene and Late Quaternary. Glacial features in the range were first recognized and described by early explorers (Gilbert, 1875; Simpson; 1876 and Russell, 1884) and subsequent authors have continued to substantiate and elaborate on earlier reports (Heald, 1956; Kramer, 1962; Currey, 1969; Peigat, 1980; Osborn and Bevis, 2001). Since this early reconnaissance work, few studies have focused on the Late Quaternary evolution and paleoclimatic implications of glacial and periglacial landforms in the Great Basin (Wayne, 1983; Osborn, 1989; Bevis, 1995; and Osborn and Bevis, 2001). However, due to the vast number of glacial deposits and ease of accessibility, the mountain ranges along the margin of the Great Basin have been extensively studied (Ray, 1940; Richmond, 1948; Moss, 1949; Richmond, 1965; Shroba, 1977; Burke and Birkeland, 1983; Birman, 1984; Dohrenwend, 1984; Mahaney, 1984; Mahaney et al., 1984; Richmond, 1986; Richmond and Fullerton, 1986; Richmond, 1986; Zielinski, 1987; and Davis, 1988).

Blackwelder (1931) conducted the first preliminary investigation on glaciation in the Great Basin and surrounding regions, describing glacial features from many mountain ranges throughout the Sierra Nevada and the Great Basin. He recognized four distinct glacial advances in the Sierra Nevada and established the terminology used today to...
describe alpine glacial stages. In stratigraphic order, from youngest through oldest, he named these stages (1) Tioga, (2) Tahoe, (3) Sherwin and (4) McGee. He based his interpretations on stratigraphic relationships between moraines in similar valleys, extent of talus cone formation, cirque morphology, modification of the valley by axial streams, weathering features, preservation of small-scale glacial features, and stream terrace formation. Subsequent workers have since refined the glacial chronology of the Sierra Nevada and established similar, tentatively correlative chronologies for the Lake Lahonton Basin, Wind River Range, and Rocky Mountains (Fig. 1.2) (Birman, 1964; Morrison and Frye, 1965; Mahaney, 1972; Miller and Birkeland, 1974; Mahaney, 1984; Mahaney et al., 1984; Richmond, 1986; Richmond and Fullerton, 1986; and Richmond, 1986).

Following the guidelines established by Blackwelder, Sharpe (1938) initiated the first detailed study of glacial features in the Great Basin. He described two discrete intervals of glaciation in the Ruby Mountains - East Humbolt Range: (1) the Lamoille and (2) Angel Lake advances. He correlated the Angel Lake advance with the Tioga stage and the Lamoille advance with the Tahoe stage, suggesting that while they were indeed two separate glaciations, they represented early and late Wisconsinan deposits. However, Wayne (1984) states there is a marked discrepancy between surficial weathering features of the two deposits, suggesting the Angel Lake advance is Wisconsinan in age while the Lamoille advance is Illinoian. Piegat (1980) conducted a preliminary investigation on the interior ranges of the Great Basin. Using aerial photographs and limited fieldwork, he defined a coarse glacial chronology for ranges in south central Nevada including the Snake Range, the adjacent Shell Creek Range, and the Toiyabe Range (located in central
<table>
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<tr>
<th>EPOCH</th>
<th>Sierra Nevada</th>
<th>Interior Great Basin</th>
<th>Lake Lahontan Basin</th>
<th>Wind River Range</th>
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<td>Tahoe</td>
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<td>Lamoille</td>
<td>Eetza-Churchill</td>
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Figure 1.2: Approximate time stratigraphic correlation diagram between glacial deposits of the Great Basin and surrounding areas in the Western United States (modified from Piegat, 1980).
Nevada at approximately the same latitude as the Snake Range). The mapping completed in this initial reconnaissance defined approximate boundaries between Lamoille, Angel Lake and Neoglacial surficial deposits. One exception to the lack of data concerning interior Great Basin glaciation is the Toiyabe Range (Ferguson and Cathcart, 1954; Stewart and McKee, 1977; and Piegat, 1980). The glacial history of this range has been formally investigated and the results suggest there were two discrete glacial advances likely correlative with Tioga-Angel Lake and Tahoe-Lamoille units (Osborn, 1989). However, the paleoclimatic significance of these deposits were not discussed, they were only correlated with regional glacial events. Osborn and Bevis (2001) provide a detailed description of the spatial distribution of Late Quaternary glaciation in ranges within the Great Basin and a paper describing the paleoclimate of each range is in progress (pg. 137).

Similarly, limited research has been conducted on the paleoclimatic significance of periglacial features in the Great Basin. Wayne (1984) describes relict rock glaciers, sorted circles, debris islands, solifluction lobes and sorted stripes in the Ruby, Schell Creek, and Snake ranges. While Currey (1969) and Osborn and Bevis, (2001) discuss the distribution of rock glaciers in numerous ranges throughout the Great Basin, including the Snake Range, and the surrounding regions. However, considerable research has been conducted on periglacial deposits located in the arid southwestern United States (Smith, 1936; Richmond, 1962; Blagbrough and Breed, 1967; Blagbrough and Farkas, 1968; Blagbrough and Breed, 1969; Galloway, 1970; Blagbrough, 1971; Barsch and Updike, 1971; Péwé, 1975; Blagbrough and Breed, 1976; Péwé and Updike, 1976; Blagbrough,
Glacial and periglacial landforms present on mountains within and adjacent to the Great Basin provide insight into the climatic conditions that prevailed throughout this region during the Late Quaternary. Alpine glaciers and permafrost are sensitive indicators of fluctuating temperature and precipitation (Péwé, 1969; Flint, 1971; Barsch, 1978; Brakenridge, 1978; Hagedorn, 1984; Wayne, 1984; Zhang, 1988; Murray, 1990; Jakob, 1992; Clark et al., 1994; French, 1996; Konrad et al., 1999; Dramis et al., 2001; Davis, 2001; Frauenfelder et al., 2001). Therefore, by obtaining understanding how certain climatic, geologic and topographic factors influence the formation and preservation of relict glacial and periglacial landforms, we can use these features to evaluate future climate changes and reconstruct Late Quaternary landscapes and paleoclimatic conditions.

Such factors include (1) the mass of a mountain range, (2) the trend of a range, (3) the proximity to and availability of moist air masses, (4) lithology, (5) slope, aspect and curvature of the landscape, (6) latitude, and (7) elevation. The mass of a mountain range above the zone of snow accumulation affects orographic precipitation and controls how much snow will accumulate on the lee slope. The greater the mass of the range the greater the chance for snow and thus an increased change that glaciers will develop (Flint, 1971). The trend of the mountain range also affects orographic precipitation; ranges that are perpendicular to moisture laden storm tracks, (i.e. – the jetstream), create a more significant barrier to this moisture. Ranges that are perpendicular to the jetstream also provide greater surface area for the accumulation of perennial snow and ice. One of the most important factors is the availability and proximity of moisture-laden air masses.
Regardless of the temperature conditions, an increase in effective moisture supply is required to develop substantial alpine glaciers in a region with such low humidity as the Great Basin. This doesn’t necessarily imply precipitation significantly increased, just that either the frequency or intensity of storm tracks increased as the jetstream was progressively depressed during the Last Glacial Maximum (LGM).

Lithology can influence curvature (i.e. – whether a surface is concave or convex) and the development of niches and slot canyons that serve as accumulation zones. Lithology can also influence the type and intensity of frost/heave sorting and thereby influence the type and distribution of permafrost (Péwé, 1969; Evans, 1994; French, 1996; and Davis, 2000). Slope can influence zones of accumulation by controlling whether perennial snow develops or whether it avalanches and accumulates at a much lower elevation (possibly below the equilibrium line [ELA] altitude of a range) (Flint, 1971 and Sharp, 1938). Avalanching may also funnel snow into a “protected” area where it survives the winter months and develops into perennial ice. Curvature also influences whether snow accumulates or not. For example, the ridgeline of most mountain crests is convex and therefore an unlikely zone of accumulation. However, slopes and valleys below the ridgeline are often convex and provide suitable zones of accumulation where nivation hollows and subsequently perennial ice can develop under appropriate climate conditions. Slope also influences which types of periglacial features develop (i.e. – solifluction lobes on dipping slopes versus patterned ground on flat level surfaces). Aspect is also of considerable importance because it influences the potential solar radiation the landscape receives throughout the year (Flint, 1971). North-northeast facing slopes provide optimum localities for the development of alpine glaciers because these slope on average, receive the least amount of solar radiation during any given year.
The aspect of ranges in the Great Basin is also important because Sharp (1938, pg. 313) identified a “wind shadow”. Although the west facing slopes of most ranges in the Great Basin receive more precipitation, snow that does manage to accumulation on ridges is blown into concave surfaces (niches, valley, hollows, etc) on the leeward side of the range. Similar to aspect, latitude influences the effectiveness of solar radiation by controlling solar duration (i.e. – length of day) and angle of incidence. Both glacial and periglacial landforms are sensitive to changes in solar radiation (Blackwelder, 1931; 1934; Flint, 1971; Funk and Hoelzle, 1992; French, 1996; Benn and Evans, 1998; Frauenfelder and Kääb, 2000), which explains the decrease in frequency and extent of glaciation from the poles to the equator (Flint, 1971). Finally, altitude directly influences precipitation, solar radiation and depressed temperatures that facilitate the development of perennial distribution of perennial snow and ice.

Specific factors that affect the development of permafrost and thus relict periglacial features include: (1) frequency of freeze-thaw cycles, (2) depth of frost penetration, (3) amount of thaw and percentage of water loss, (4) presence of water near the surface, (5) depth of snow, and (6) vegetation cover (Péwé, 1969; Harris, 1981; Evans, 1994; French, 1996; and Davis, 2000). The frequency of freeze-thaw cycles and depth of frost penetration influence the degree of sorting and development for some permafrost features. The seasonal water loss associated with summer thaw can influence whether sporadic or discontinuous permafrost develops. The presence of water at the surface can influence summer preservation of permafrost by either reflecting solar radiation (i.e. – increased albedo) or if abundant water is present, by radiating absorbed solar energy into the ground. The depth of snow cover (>50 cm) during the winter months has been shown to significantly affect the development of permafrost (Harris, 1981 and French,
1996) by insulating the ground and preventing the development of ground ice and thus permafrost.

Finally, vegetation is the most complex variable because numerous interactions take place between vegetation and the subsurface that are still poorly understood. Vegetation also serves as an insulating barrier to solar radiation and the insulating effect is tree and plant specific (Viereck, 1965 and French, 1996).

Understanding the conditions required for permafrost features to develop also provides further insight into the genesis and preservation of periglacial landforms. In addition, by understanding landform genesis we can begin to evaluate whether modern features still retain interstitial and/or glacial ice. Rock glaciers in particular are potential ancillary sources of water, something of particular interest in the southwestern United States (Konrad et al., 1998; Burger et al., 1999; Degenhardt and Giardino, 2003; Degenhardt et al., 2003; and Whalley and Azizi, 2003). Currey (1969), Piegat (1980) and Osborn and Bevis (2001) describe the occurrence of numerous rock glaciers throughout the Great Basin, including the Snake Range. These landforms are potentially sequestering interstitial or glacial ice that may be released and utilized as global climate continues to warm. Unfortunately, we still have a incomplete inventory of the relict rock glaciers in the Great Basin and a poorer understanding of their ice content because of the lack of research being conducted in the interior Great Basin on periglacial landforms. However, with the current demand for water in the southwestern United States and the projected mean global temperature warming of 1.5-5°C (Watson et al., 1990; Wilson and Mitchell, 1987; Schlesinger and Zhao, 1989; Knox, 1991; and Mitchell et al., 1995) these potential water reserves will certainly generate more interest.
The goals of this project were to (1) provide a better understanding of the surficial deposits associated with glacially derived landscapes in the Snake Range, (2) evaluate the genesis and spatial distribution of glacial and periglacial deposits in the Snake Range, (3) derive paleoclimatic estimates from these deposits using field evidence and computer modeling, and (4) investigate the micro-characteristics of glacially derived clasts and debris pertinent to alpine glacier studies. Chapter 2 describes the geomorphology and paleoclimatic implications of glacial and periglacial deposits in the Snake Range and evaluates the hypotheses that (1) solar radiation is an applicable tool for predicting the distribution of permafrost and (2) the Snake Range records a paleoclimate signature similar to adjacent ranges in Nevada and Utah (Bevis, 1995). Chapter 3 describes the geomorphology of the Lehman rock glacier (LRG) in the southern Snake Range and evaluates the hypothesis that (1) the LRG is composed of three distinct lobes and evolved in response to the retreating Lehman Glacier. Chapter 4 presents results from computer modeling that attempt to predict Late Quaternary temperature and precipitation conditions in the Snake Range. My hypothesis was that using a snow and ice model, I could predict temperature changes required to develop the moraines identified in the field. Chapter 5 describes the micromorphology of glacially affected clasts and surfaces from the southwestern United States and the Arctic. This study tested the hypothesis that distinct micro-features develop on during transport or scouring by glaciers.

Each chapter addresses one or more paleoenvironmental conditions required to initiate, develop, and support glacial and periglacial environments. The culminating papers describe the genesis and distribution of landforms, suggest possible paleoclimate scenarios, and suggest that glacial and periglacial processes were pervasive during the LGM and the Late Quaternary.
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CHAPTER 2

THE PALEOCLIMATIC SIGNIFICANCE OF RELICT GLACIAL AND PERIGLACIAL LANDFORMS, SNAKE RANGE, NEVADA

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Introduction

Previous researchers have described the existence and general characteristics of glacial and periglacial landforms in the interior Great Basin (Blackwelder, 1931, 1934; Sharp, 1938; Currey, 1969; Waite, 1974; Piegat, 1980; Wayne, 1983, 1984; Osborn, 1990; Osborn and Bevis, 2001). Osborn (1990) and Osborn and Bevis (2001) have discussed the glacial history of the Snake Range. However, only preliminary studies have focused on glacial and periglacial deposits in the Snake Range in east-central Nevada (Heald, 1956; Kramer, 1962; Piegat, 1980; Wayne, 1983; Osborn, 1990; and Osborn and Bevis, 2001). This study presents a detailed description and inventory of relict rock glaciers and periglacial features in the Snake Range, and an interpretation of their paleoclimatic significance.


Continuous permafrost is perennally frozen ground that occurs in regions with a soil temperature of at least 0°C for two consecutive years (Péwé, 1983 and Davis, 2000). Discontinuous permafrost is also perennially frozen ground occurring near the 0°C isotherm, however it is disrupted by talik, layers of unfrozen soil between the permafrost table and the active layer (Péwé, 1983 and Davis, 2000). A number of periglacial features identified in the southwestern United States are indicative of discontinuous alpine permafrost (Smith, 1936; Richmond, 1962; Blagbrough and Breed, 1967; Blagbrough and Farkas, 1968; Blagbrough and Breed, 1969; Galloway, 1970; Blagbrough, 1971; Barsch and Updike, 1971; Péwé, 1975; Péwé and Updike, 1976;
Blagbrough, 1976; Wayne, 1984; Nicholas, 1994; and Nicholas and Butler, 1996).

Estimates for the lower limit of discontinuous alpine permafrost are variable and range from -8.5°C in the Alps to 0.5°C on the Qinghai – Xizang Plateau, China (Haberli, 1978; Harris, 1981 a, b, c; Wayne, 1983; Guodong, 1983; Frauenfelder and Kääb, 2000; and Frauenfelder et al., 2001). The high variability in temperature estimates is a function of localized alpine microclimates governed by aspect, slope, solar insolation, albedo, and snowcover (Harris, 1981 a, b, c; Barry, 1981; and Greenstein, 1983). Although many periglacial features suggest the presence of discontinuous permafrost, rock glaciers have the greatest potential for estimating the lower limit of discontinuous alpine permafrost and therefore, the location of the 0°C isotherm (Pewé, 1969; Barsch, 1978; Jakob, 1992; and Frauenfelder et al., 2001). The paleoclimatic significance of rock glaciers has only been investigated in a few areas, even though they are intimately related to permafrost and limited snowfall, implying that they have considerable potential for paleoclimatic studies (Kerschner, 1978, 1980 and 1985; Buchenauer, 1990 and Humlum, 1998a, 1998b, 1999a, 1999b).

The use of rock glaciers in paleoclimate studies are limited because we require a better understanding of the climate typical of rock glaciers, however as climate data becomes more available from sites with active rock glaciers, these studies should become more prevalent. However, the presence of periglacial landforms indicates that at the very least, discontinuous alpine permafrost existed throughout the Snake Range during the Neoglacial and the Last Glacial Maximum (LGM; Péwé, 1983). All of the features identified in this study are inactive, indicating the 0°C isotherm was lowered during the Neoglacial and LGM.
Numerous methods have been employed to estimate temperatures for the southwestern United States during the last Full Glacial (~20 ka). These studies suggest a range of temperature depression from 2.8°C to 10°C (Broecker and Orr, 1958; Snyder and Langbein, 1962; Brackenridge, 1978; Mifflin and Wheat, 1979; Dohrenwend, 1984; Benson, 1986; Benson and Thompson, 1987; and Bevis, 1995). This study calculates temperature depression estimates based on the elevation and distribution of relict permafrost landforms and differences in modern mean annual air temperature (MAAT) and calculated MAAT for Neoglacial and Full Glacial environments. These estimates are compared with temperature depression estimates garnered from equilibrium line altitudes (ELA) of reconstructed glaciers. ELAs are estimated using three techniques and two sets of regional lapse rates to calculate minimum and maximum temperature estimates.

Study Area

The Snake Range is located in east-central Nevada within the boundaries of the Basin and Range physiographic province. The entire region experienced Cenozoic extension resulting in extensive normal faulting. Along these roughly north-south-trending faults, mountains were uplifted and valleys down-dropped, producing horst and graben topography. The Snake Range is an uplifted horst bounded to the east by Snake Valley and to the west by Spring Valley (Fig. 2.1). Topographically isolated regions such as the Snake Range are found throughout the Basin and Range province and produce drastically different micro-climates compared to the surrounding low lying valleys. The climate is strongly influenced by both elevation and aspect.
Figure 2.1: Study area map depicting the location of landmarks and permafrost features. CT = cryoplanation terrace, SL = solifluction lobes, PR = protalus rampart, G = garlands, and PG = patterned ground.
Average mean annual air temperatures within the Snake Range vary from -8°C to 30°C, and average annual precipitation from the valleys to the high peaks region varies from less than 12 cm to greater than 50 cm (Houghton, 1969). The western Great Basin is characterized by precipitation patterns that deliver a significant portion of mean annual precipitation in the monsoon season during July and August (Houghton, 1969). Previous studies have identified the “Great Basin High,” a thermally derived air mass that remains stationary during much of the winter season and forces westerly storms north of 40°N latitude (Houghton, 1969 and Mitchell, 1976). The Great Basin High is ephemeral and is often depressed during the winter, providing a storm path to the southern portion of the Great Basin. Much of the precipitation delivered to the interior regions of the Great Basin occurs during these periods dominated by the “Pacific component” (Houghton, 1969). However, most of the precipitation received by the Snake Range occurs during April to June and October to November when successive low-pressure cells dominate the regional climate under the “Continental component” (Houghton, 1969).

The Snake Range is composed of early Paleozoic Prospect Mountain quartzite, Pole Canyon limestone, and Pioche shale intruded by the Snake Creek–Williams Canyon pluton that is Jurassic age (~160 Ma; Drewes, 1958; Whitebread, 1969; Lee et al., 1970; Lee and Van Loenen, 1971; Lee et al., 1981; Lee and Christiansen, 1983 a, b; Lee et al., 1986; Miller, 1995 and Miller, 1999). The Snake Range contains an excellent example of a Basin and Range metamorphic core complex décollement. The décollement is thought to have accommodated 12–15 km of rapid slip during the Miocene (17 Ma; Lewis et al., 1999 and Miller et al., 1999), and an older episode of faulting is thought to have occurred during the late Eocene – Oligocene (Miller et al., 1999).
Piecat (1980), Osborn (1990), and Osborn and Bevis (2001) summarize the glacial geology in the Snake Range and correlated Late Wisconsin deposits with the Angel Lake advance (AL) and Illinoian deposits with the Lamoille (LAM) advance. The AL and LAM terminology was defined in the Ruby Mountains by Sharp (1938) and later applied to the East Humbolt Range by Wayne (1984) and throughout the western Great Basin by Piecat (1980) and Osborn and Bevis (2001). Angel Lake and Lamoille moraines have also been correlated with Blackwelder’s (1934) Tioga and Tahoe advances in the Sierra Nevada. Sharpe (1938) and Wayne (1984) provide excellent descriptions of the characteristics used to differentiate between Angel Lake and Lamoille glacial moraines.

Methods

We used aerial photography interpretation and field mapping during the spring and summer of 2001 and 2002 to locate and identify glacial and periglacial landforms. We obtained color aerial photographs (1:24,000 and 1:12,000 scale) from the United States Department of Agriculture and black and white aerial photographs (1:48,000) from the United States Geological Survey for the Snake Range. We mapped landforms on digital orthophotos using a Geographic Information System (GIS) and later compared with results from fieldwork. Rock glaciers are described and classified using the taxonomy of Barsh (1996) and Corte (1987). We describe and measure the morphometric parameters of each rock glacier using a GIS following guidelines described by Barsh, (1996).

Using a 30-meter digital terrain model (DTM), modern climate data and Visual Basic programming we calculate daily mean annual air temperatures (MAAT) for every grid cell in the DTM and freezing and thawing indices for five rock glaciers in the Snake Range. We also estimated MAAT using a thirty-year temperature record from the Lehman Caves climate station and lapse rates defined by Greenstein (1983), Dohrenwend (1984) and Wolfe (1992) (Table 2.1).

<table>
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<tr>
<th>Source</th>
<th>Altitude (m)</th>
<th>Lapse Rate</th>
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</thead>
<tbody>
<tr>
<td>Greenstein, 1983</td>
<td>&gt; 2,600</td>
<td>5.0 °C/Km</td>
</tr>
<tr>
<td>Wolfe, 1992</td>
<td>2,400 – 2,600</td>
<td>3.7±0.2°C/Km</td>
</tr>
<tr>
<td></td>
<td>2,200 – 2,400</td>
<td>2.6°C/Km</td>
</tr>
<tr>
<td></td>
<td>2,000 – 2,200</td>
<td>2.4±0.4°C/Km</td>
</tr>
<tr>
<td></td>
<td>1,800 – 2,000</td>
<td>2.5±0.8°C/Km</td>
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<tr>
<td></td>
<td>1,600 – 1,800</td>
<td>2.2±0.4°C/Km</td>
</tr>
<tr>
<td>Dohrenwend, 1984</td>
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</tr>
<tr>
<td></td>
<td>&lt; 2,000</td>
<td>7.6°C/Km</td>
</tr>
</tbody>
</table>

Table 2.1: Summary of lapse rates used to calculate modern and past MAAT for paleotemperature depression estimates.

The Lehman Caves climate station is located on the east side of the southern Snake Range at 2073 meters above sea level (masl) and has a MAAT of 9.59°C. Lapse rates were not corrected for the air drainage effect described by Barry (1981) and Harris (1982) because of the lack of climate stations in the study area. Freezing and thawing indices were calculated for each rock glacier using daily MAAT values and plotted versus the lower limits of continuous, discontinuous, and sporadic permafrost calculated by Harris (1981a, b, c). By assuming the terminus of each rock glacier represents the paleo-distribution of the lower limit of discontinuous permafrost, paleotemperatures can be
calculated for each rock glacier by depressing the modern lower limit of discontinuous permafrost and measuring the temperature change.

Paleotemperatures are also calculated from rock glacier termini using a digital terrain model, potential direct radiation values and MAAT following the methodology described by Frauenfelder et al. (2001) and Frauenfelder and Kääb (2000). Paleotemperatures were estimated by calculating the current MAAT \( T_p \) at the terminus of each rock glacier using modern climate data and lapse rates. Potential direct radiation values \( I_p(x,y,z) \) were calculated for the winter and summer solstice using ArcView GIS and The Solar Analyst Extension, a radiation model that utilizes a similar algorithm to Funk and Hoelzle (1992). A hypothetical altitude \( H_{lim} \) for each relict rock glacier is calculated representing the elevation required for the rock glacier to be considered active using Equation (1):

\[
H_{lim} = aI_p(x,y,z) + b
\]

where \( a \) and \( b \) are constants calculated from BTS measurements in the Swiss Alps by Hoelzle and Haeberti (1995) and \( I_p \) is modern potential direct solar radiation at the terminus of each rock glacier. A hypothetical MAAT \( T_{lim} \) was calculated using the hypothetical altitude \( H_{lim} \) as a proxy for temperature using Equation 2:

\[
T_{lim} = \left(C - H_{lim}\right)\frac{\delta T}{\delta h}
\]

where \( C \) is the elevation of the regional 0° isotherm and \( \delta T/\delta h \) is the regional lapse rate.

These calculations provide MAAT under modern conditions and paleo-MAAT estimates that can be used to calculate temperature change using Equation 3.

\[
\Delta T = T_p - T_{lim}
\]

The difference \( \Delta T \) between \( T_{lim} \) and \( T_p \) represents the temperature depression from the time each rock glacier developed and modern climate.
The spatial distribution of periglacial features is also used to estimate MAAT depression in the Snake Range during the Last Glacial Maximum (LGM). Using the modern MAAT of 9.60°C and regional lapse rates (Greeenstein, 1983; Dohrenwend, 1984; and Wolfe, 1992; Table 2.1), minimum and maximum estimates were calculated for full glacial cooling in the Snake Range.

Paleotemperature estimates are calculated from modern and former equilibrium line altitudes (ELA). ELA values are calculated using the spatial extent of Angel Lake and Lamoille glaciers in the Snake Range following guidelines described by Meier and Post (1962), Porter (1975), Dohrenwend (1984), Meierding (1982), Furbish and Andrews (1984), Tomes et al. (1993), Benn and Gemmell (1997), and Benn and Evans (2001).

First order estimates of Angel Lake and Lamoille ELAs were calculated using the toe to headwall altitude ratio (THAR; Charlesworth, 1957 and Richmond, 1965). The highest probable extent of glacial ice and lowest altitude of each terminal moraine were calculated for Angel Lake and Lamoille glacial lobes. THAR ratios of 0.35 and 0.40 are assumed to most accurately represent paleo–ELAs because they produce the lowest root mean square error (RMSE; Meierding, 1982). However, ELAs calculated using a THAR ratio of 0.50 produce the most realistic values in the Snake Range when compared to the spatial extent of glaciation.

ELAs were also calculated using the accumulation area ratio (AAR) method. This methodology is more sensitive than the THAR technique because it assumes a fixed proportion of the total glacier is located in the zone of accumulation (Meierding, 1982; Locke, 1990; and Benn and Evans, 2001). This technique calculates former ELA values by identifying the altitude on a glacier that corresponds to a specified accumulation area (0.50, 0.55, 0.45, 0.40, etc). This technique, using an AAR of 0.65, most accurately
(±100 m) represents the ELA of former glaciers based on RMSE error analyses (Meierding, 1982 and Gillespie, 1991).

However, the AAR technique does not evaluate the hypsometry of former glaciers and can produce erroneous ELA values. Therefore, ELAs were also calculated using the balance ratio (BR) method developed by Furbish and Andrews (1984) and automated by Benn and Gemmell (1997). The BR method incorporates the assumption that net accumulation must exactly balance net ablation below the ELA using equation 4:

\[ d_b A_b = d_c A_c \]  

where \( d_b \) equals the average net annual ablation in the ablation zone, \( A_b \) is the area of the ablation zone, \( d_c \) is the average net annual accumulation in the accumulation zone and \( A_c \) is the area of the accumulation zone (Furbish and Andrews, 1984). The ablation and accumulation gradients are assumed to be linear so that \( d_b \) and \( d_c \) are in equilibrium at the area-weighted mean altitudes of the ablation area \( (z_b) \) and accumulation area \( (z_c) \). A steady-state ELA is defined by the altitude that balances equation 5.

\[ b_{nb}/b_{nc} = z_c/A_c/z_b/A_b \]  

where \( b_{nb} \) is the mass balance gradient in the ablation zone and \( b_{nc} \) is the mass balance gradient in the accumulation zone. Typically, small ablation zones with respect to glacial area characterize glaciers exhibiting high balance ratios and glaciers with lower balance ratios require larger ablation zones for equilibrium. BR values were calculated using the spreadsheet model provided by Benn and Gemmell (1997).

First and second order local al and global polynomial surfaces are calculated for the Snake Range using values calculated from each method using ArcGIS. Contouring these surfaces produced ELA distribution maps for each technique and allowed for the calculation of RMSE and correlation statistics (Table 2.2; Meierding, 1982).
Results

Permafrost Features

Relict permafrost features identified in the Snake Range include rock glaciers, cryoplanation terraces, protalus ramparts, stone polygons, stone circles, solifluction lobes, and garlands (Fig. 2.1). The location and characteristics of these landforms are summarized in Table 2.2. The most common periglacial features in the Snake Range are stone–banked solifluction lobes (8 sites), followed by stone polygons and circles (6 sites), rock glaciers (5 sites), cryoplanation terraces (5 sites), garlands (3 sites), and protalus ramparts (3 sites). No preserved organic material has been discovered in any of the features that would allow for radiocarbon dating. We assume these features developed during the last glacial maximum, and during early deglaciation with the exception of rock glaciers. Tephrachronology data indicates the presence of Mono Crater ash in the lower two segments of the Lehman Cirque rock glaciers, indicating that these features must be younger than 1200 years (Osborn, 1990).

Rock glaciers

The Snake Range contains 5 alpine rock glaciers located in Lehman Cirque, North Fork Baker Cirque, Teresa Cirque, Big Canyon, and on the northwest slope of Jeff Davis Cirque (Fig. 2.1). All five rock glaciers are located on or below steep cliffs composed of Prospect Mountain Quartzite and are characterized by high rates of erosion and talus production.

The Lehman rock glacier is approximately 900 m long. It is located in a northeast-facing cirque composed entirely of Prospect Mountain Quartzite and is directly connected.
<table>
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<th>Feature</th>
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<th>Lithology***</th>
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* WPR = Wheeler Peak Ridge, TC = Teresa Cirque, NFBC = North Fork Baker Cirque, MMT = Mt Moriah Table, MM = Mt Moriah, LP = Lincoln Peak, LC = Lehman Cirque, JP = Johnson Pass, JDR = Jeff Davis Ridge, JDC = Jeff Davis Cirque, BM = Bald Mountain, and BR = Baker Ridge.

** Stone circles and polygon size measures the diameter and garlands and solifluction lobes measure length.

*** PMQ = Prospect Mountain Quartzite, SCWP = Snake Creek-Williams Pluton, and PCL = Pole Canyon Limestone.

Table 2.2: Summary of general characteristics and calculated temperature change at each site since the last Full Glacial.
to the source area (Fig. 2.2). The rock glacier is a complex, tongue-shaped feature dominated by three distinct lobes composed of large (0.5 to 3.0 m) blocky quartzite clasts with a minor component of fine-grained sediment. Deep, arcuate furrows separate each lobe from one another. The upper lobe (LUL) of the rock glacier has a rooting elevation of approximately 3667 masl, and the lower lobe (LLL) terminates at approximately 3272 masl. The surface of the rock glacier exhibits well-developed surface relief ranging from 2.0 to 3.0 m in the form of prominent furrows and ridges that are oriented perpendicular to down-valley motion.

The upper lobe (LUL) is a convex landform approximately 434 meters long, 206 meters wide and exhibits well-developed arcuate furrows and ridges that range in amplitude from 0.5 – 3.0 meters. The frontal and lateral margins of the lobe are composed of light gray sediment and talus. They are steep (~34 – 36°), and capped by a 0.5 m thick sequence of dark blocky talus. The frontal margin lacks boulder aprons, and sparse vegetation is found growing on the lowermost surface.

The middle lobe (LML) exhibits a moderately convex surface that is approximately 265 m long and 178 m wide. This lobe also exhibits moderately developed one-to-two-meter deep arcuate furrows and ridges. The frontal margin is composed of fine-grained sediment and talus overlain by a darker sequence of blocky talus that is approximately one meter thick. The frontal slopes are steep (~32 – 36°), with small poorly developed boulder aprons are present at the base of the steep frontal snout. This lobe contains significantly more vegetation than the upper lobe. The lateral margins of the lobe are difficult to distinguish because of talus input and subsequent coalescence of the rock glacier and talus cones.
Figure 2.2: Morphometric map of the Lehman rock glacier. The inset photo of the Lehman rock glacier complex was taken approximately 500 meters downvalley on the Lehman-Bristlecone trail.
The lower lobe (LLL) is a weakly convex landform approximately 195 m long and 100 m wide, lacking well developed surface furrows and ridges. The frontal margin is dominantly fine-grained sediment and talus, overlain by a heavily weathered layer of blocky talus. The frontal slopes are steep (\(\sim 30 - 33^\circ\)). Numerous small bushes and flowers exist on the slope, and well-developed boulder aprons and evidence for small rockslides are present at the base of the steep frontal snout.

The North Fork Baker Creek rock glacier (NFBC) is a moderately convex tongue-shaped landform approximately 673 m long, 88 m wide. It is located in an east-to-northeast facing cirque composed of Prospect Mountain Quartzite, and it is directly connected to the source area (Fig. 2.3). This is a complex landform composed of multiple lobes that have evolved into the modern, tongue-shaped feature. Although there are multiple lobes in this rock glacier complex, it is difficult to separate them into distinct rock glaciers because of high rates of talus input from the surrounding cirque walls. This rock glacier is located below a steep, north-facing ridge (approximately 50 to greater than 60 degrees) that provides significant shading during the summer months.

The surface of the rock glacier is dominated by blocky quartzite talus (0.5 – 1.5 m) and is characterized by well-developed surface relief (0.5 – 2.5 m) in the form of prominent arcuate shaped furrows and ridges. The rooting zone for the upper segment of the rock glacier is located at approximately 3397 masl. The true rooting zone may be located at a higher elevation, but this is not discernible in the field or from aerial photos because of thick talus cover. The lowermost segment terminates at ~ 3261 masl. The frontal and lateral margins of the rock glacier are steep (\(\sim 32 - 34^\circ\)), with reddish-gray sediment overlain by a 0.5 m layer of dark, blocky talus. The margins of the rock glacier exhibit moderately developed boulder aprons and contain sparse vegetation.
Figure 2.3: (a) Morphometric map of the North Fork Baker rock glacier located south of the Lehman rock glacier and (b) the location where the field photo was taken is designated with a star.
The Teresa rock glacier (TR) is a single, lobate-shaped convex upward landform approximately 147 m long and 225 m wide. It is located in a small east-to-northeast facing cirque composed of Prospect Mountain Quartzite, and it is detached from the source area headwall (Fig. 2.4). Although this is a relatively small feature, the steep frontal snout (~33 – 35°) and moderately developed furrows and ridges (1.0 – 1.5 m) suggest that this is a rock glacier, and not a protalus rampart. The uppermost portion of the rock glacier is located at approximately 3319 masl, and the lowermost portion is located at approximately 3260 masl. Well-developed boulder aprons are present at the base of the frontal snout, and evidence of small debris slides exists along the upper portion of the snout. Numerous shrubs and grasses are growing on the frontal slope.

The North Slope Jeff Davis rock glacier (NSJD) is a single, weakly convex-upward, tongue-shaped landform approximately 230 m long and 119 m wide. It is found on a north-facing slope of Jeff Davis Peak that is composed of Prospect Mountain Quartzite (Fig. 2.5). This landform is the only identified rock glacier in the Snake Range that is not located in a cirque or below a steep-faced ridge. This rock glacier has a steep frontal snout (~28 – 33°) and weakly developed furrows and ridges (0.5 – 1.0 m) suggesting that is a rock glacier and not a protalus rampart. The uppermost portion of the rock glacier is located at approximately 3402 masl and the lowermost portion is located at approximately 3318 masl. Moderately developed boulder aprons are present at the base of the frontal snout, and sparse vegetation is found growing on the frontal slope.

The Big Canyon rock glacier (BCR) is a single, weakly convex-upward, lobate-shaped landform approximately 172 m long and 119 m wide. It is located below a west-facing slope on Mt Moriah, which is composed of Prospect Mountain Quartzite (Fig.
Figure 2.4: (a) A morphometric map of Teresa rock glacier that is located to the north of Lehman rock glacier and (b) the red star on the aerial photograph is the approximate location where the photograph was taken.
Figure 2.5: (a) A morphometric map of the North Slope Jeff Davis rock glacier and (b) a photo taken from the Wheeler Peak Scenic Drive facing southwest.
Figure 2.6: (a) A morphometric map of the Big Canyon rock glacier located north of Mt. Moriah and (b) the red star indicates the approximate location where the field photo was taken.
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The ellipse shows the prediction interval for a single new observation, given the parameter estimates for the bivariate distribution computed from the data, and the given n.

Regression Results: $r^2 = 0.9991; \ y = 3330.16349 - 198.771969x$

Figure 2.7: Rock glacier terminus elevation versus MAAT calculated from climate data.

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2.6). The landform has a steep frontal snout (~28 – 33°) and weakly developed furrows and ridges (0.5 – 1.0 m) suggesting that this is a rock glacier, and not a protalus rampart. The uppermost portion of the rock glacier is located at approximately 3402 masl and the lowermost portion is located at approximately 3318 masl. Moderately developed boulder aprons are present at the base of the frontal snout, and sparse vegetation is found growing on the frontal slope.

**Paleoclimatic Significance of Snake Range Rock Glaciers**

Rock glaciers have been used to estimate temperature change under the assumption that they represent the lower limit of discontinuous permafrost (Barsch, 1978; Jakob, 1992; and Frauenfelder et al., 2001). Modern climate conditions were established to quantify temperature change since the development of each rock glacier in the Snake Range. The MAAT for the lower limit of each rock glacier is plotted against the regional MAAT and elevation of the Snake Range in Figure 2.7. Freezing and thawing indices calculated for each rock glacier are plotted against the freezing and thawing indices for the limits of continuous, discontinuous, and sporadic permafrost (Fig. 2.8; Harris, 1981a, b, c). Assuming rock glaciers represent the lower limit of discontinuous permafrost, they should plot above the discontinuous permafrost isotherm. However, under modern climate conditions every rock glacier falls below this isotherm, indicating the region has cooled since they developed or the assumption that they represent the lower most limit of discontinuous permafrost is invalid. The difference between modern freezing and thawing indices and the lower limit of discontinuous permafrost for each rock glacier should represent the temperature change between modern climate and the climate under which each feature developed (Table 2.3). Temperature depression estimates since the development of Neoglacial rock glaciers range from –0.25°C for the
Figure 2.8: Freezing and thawing indices for 7 rockglaciers in Snake Range, Nevada, plotted with respect to the lower limits of the continuous, discontinous, and sporadic permafrost zones (After Harris, 1981,a, b, c and Greenstein, 1983)
lower lobe of the Lehman rock glacier to \(-1.0\)°C for the Big Canyon rock glacier, and the estimates have a mean of \(-0.61\)°C.

GIS-based modeling was also used to calculate Neoglacial temperature change estimates based on (a) the elevation of the lower limit of each rock glacier, (b) summer and winter radiation values, and (c) modern MAAT following the methodology of Frauenfelder and Kääb (2000) and Frauenfelder et al. (2001). Snake Range rock glaciers are located in an altitudinal range of 3144 to 3336 masl with aspects ranging from 8° to 280° and they are subjected to a broad range of winter and summer solar radiation values.

Temperature depression estimates since the development of Neoglacial rock glaciers, calculated using the relationship between modern potential direct solar radiation and MAAT, range from \(-1.11\)°C for the North Fork Baker Cirque rock glacier to \(-1.61\)°C for the Jeff Davis NW Slope rock glacier, with an estimated mean temperature change of \(-1.42\)°C (Table 2.3). Modern MAAT (\(T_p\)), the hypothetical temperature (\(T_{lim}\)) at the Neoglacial rock glacier front, and the calculated difference (\(\Delta T\)) for each rock glacier are plotted in Figure 2.9. The distribution of temperature change in Figure 2.10 indicates a similar relationship to that described above, with the freeze/thaw indices; the lowest temperature depression occurs at the lower lobe of the Lehman rock glacier and the greatest depression occurs at the Big Canyon rock glacier. This temperature gradient is consistent with field observations because the lower lobe of the Lehman rock glacier is presumed to be the youngest rock glacier in the Snake Range and is in close proximity to the Lehman glacier and modern ELA, while the Big Canyon rock glacier faces directly west and is approximately 192 m lower than lower lobe of the Lehman rock glacier.

The average temperature depression estimate of \(-0.61\)°C, calculated using freezing and thawing indices, is lower than previously calculated Neoglacial temperature
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Mean \( \rightarrow \) 3253.47335 -0.6142857

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Mean \( \delta T \rightarrow \) -1.423892

Figure 2.9: Comparison of modern and Neoglacial MAAT calculated for the terminal elevation of Snake Range rock glaciers.
depression estimates in the Southwest (Wayne, 1984). Information about winter snow cover is not available for the Snake Range, but these anomalously low temperature estimates may be related to a winter snow cover of greater than 50 cm. Snowcover greater than this critical thickness can produce erroneous results (Harris 1981, a, b, c).

The average temperature depression estimate of \(-1.42^\circ C\), calculated using MAAT and potential direct solar radiation, is close to the range of previously calculated Neoglacial cooling estimates of \(-1.5^\circ \) to \(2^\circ C\) (Wayne, 1984).

Other Periglacial Features

The spatial distribution of patterned ground is often controlled by lithology, clast size, precipitation, moisture conditions, and slope (Washburn, 1973; Troll, 1994; French, 1996; Davis, 2000; Benn and Evans, 2001). However, the size and elevation of relict permafrost features can be used to estimate paleotemperature and extent of permafrost during the full and late glacial conditions (Péwé, 1969; Washburn, 1980; Harris, 1981a; Péwé, 1983; and Wayne, 1983).

The site with the most extensive preservation of stone-banked solifluction lobes is the east to southeast facing slope of Jeff Davis Peak. The lobes on Jeff Davis Peak are most easily identified in the late spring when snow remains in the hollows around and between the lobes. These lobes range in length from 3 to 7 meters and occur on slopes ranging from 21 to 33° degrees. Solifluction lobes are also present on the northern slopes of Bald Mountain, Mt Moriah, Johnson Pass and Baker Ridge. These features range in length from 0.5 to 8 m on slopes ranging from 27 – 36° (Fig. 2.10).

The most extensive area of sorted polygons and stone circles is located on “The Table”, a relict upland surface on the eastern side of Mt Moriah. The large diameter of
Figure 2.10: (a) well developed solifluction lobes on a east to northeast facing slope on Jeff Davis Peak and (b) view of well developed solifluction lobes on Mt Moriah. The inset is an 1:12,000 aerial photo illustrating how extensive the lobes on Mt Moriah.
the polygons and circles on Mt. Moriah are comparable to patterned ground of Wisconsin age in the Colorado Front Range (Pewe, 1983). Sorted polygons range in diameter from 4.0 to 5.2 m, composed of tabular blocks about 0.3 to 1.0 m wide. Stone circles range in diameter from 1.3 to 1.8 m. Other smaller stone polygons and circles were found on Lincoln Peak, Johnson Pass, Baker Ridge, and on the lower lobe of the Lehman rock glacier. These features range in size from 0.3 – 3.5 m and occur in an elevation range of 3332 to 3664 masl. Patterned ground primarily occurs on cryoplanation terraces and relatively flat surfaces in Prospect Mountain Quartzite. Prospect Mountain Quartzite is the dominant lithology with the exception of stone polygons that developed on Lincoln Peak in Pole Canyon Limestone and stone circles that developed in Johnson Pass in Snake Creek-Williams Canyon granite (Fig. 2.11).

Most of these features are relatively small and it could be argued that they developed because of less intense sorting. However, the larger polygons and circles developed in competent, quartzite debris, and the smaller features developed in less competent limestone and granite parent material. This suggests that the development of these features may be a function of lithology rather than MAAT. Stone polygons and circles developed adjacent to areas affected by glaciation with the exception of small stone circles identified on the lower lobe of the Lehman rock glacier. This suggests some areas developed permafrost features during the Neoglacial. However, other than the sorted circles on the lower lobe of the Lehman rock glacier, it is impossible to differentiate between Full Glacial and Neoglacial features without absolute ages.

Five cryoplanation terraces were identified in close proximity to patterned ground. The terraces, which are located predominantly on northeast and northwest-facing slopes, most likely developed in response to nivation and long periods of freeze-thaw sorting.
Figure 2.11: The various types of patterned ground observed on The Table and on Johnson Pass. (A) Sorted circles developed in quartzite on The Table (scale = 1.5 m), (B) sorted circles developed in quartzite north of The Table (scale = 1.5 m), and (C) non-sorted circle developed in granite on Johnson Pass (scale = 1.3 m).
The largest terrace occurs on Mt Moriah as “The Table”, at approximately 3700 masl that has an area of 1,163 m². This terrace is located on the northeast slope of Mt Moriah and has little topographic protection from incoming solar radiation. Two terraces are located on the ridge between Bald Mountain and Wheeler Peak; one is located on the ridge between North Fork Baker Cirque and Baker Cirque, and another is located between Baker Cirque and Johnson Cirque in Johnson Pass (Fig. 2.12). They range in size from approximately 43 m² to 66,931 m². The distribution of the terraces is consistent with the idea that snow patches tend to accumulate away from prevailing winds and direct sunlight. In the Great Basin, prevailing winds on the crest of the ranges usually come from the west and southwest (Houghton, 1969).

Moderately developed garlands are rare in the Snake Range. This may not reflect a lack of landform preservation, but rather that the slopes on which garlands typically occur are steep and often inaccessible. So it is possible that other sites exhibit garlands that were not identified (Fig. 2.13). Garlands were identified on a south-facing slope of Jeff Davis Peak and on a south to southeast-facing slope on Baker Ridge. The three sites with garlands are all on slopes ranging from 29° to 33° in an elevation range of 3500 to 3700 masl. Garlands range in size from 1.0 to 4.0 m long and approximately 8 to 15 m wide.

Three protalus ramparts occur in two locations; two are located at the head of North Fork Baker Cirque and one is below a north-facing cliff in Teresa Cirque (Fig. 2.13). The two identified in North Fork Baker Cirque are east to northeast-trending lobate landforms, and the rampart in Teresa Cirque is a northwest-trending complex landform. These features range in size from 3 to 4 meters high and 15 to 30 m long, and they lack evidence of fine-grained sediment. The long axis of each landform is parallel to the
Figure 2.12: (a) A view of two cryoplanation terraces on Wheeler Peak ridge (looking north to northeast) and (b) a view looking east of "The Table" on Mt Moriah in the Northern Snake Range, an extensive well developed cryoplanation terrace.
Figure 2.13: (a) A view looking north to northwest of moderately developed garlands on a south-facing slope of Jeff Davis Peak and (b) a view looking east towards a protalus rampart and talus cone complex in Teresa Cirque.
source talus slope, and a well-developed nivation hollow separates each feature from the talus slope.

Discussion

Paleoclimatic Significance of Snake Range Periglacial Features

The temperature thresholds for continuous and discontinuous permafrost are discussed by Fenians (1965), Billings and Mooney (1968), Brown (1970), Brown and Péwé (1973), Washburn (1979), and Wang and French (1994). Estimated MAATs for discontinuous and permafrost range from $-1^\circ$C to $-13^\circ$C. We use the estimates provided by Billings and Mooney (1965) because they report more significant figures (e.g. $-3.3^\circ$C to $-12.4^\circ$C). Temperature depression estimates since the development of Full Glacial permafrost features were calculated using modern MAAT and regional lapse rates.

Temperature depression estimates using lapse rates from Dohrenwend (1984) range from $-5.02^\circ$C, calculated from solifluction lobes located on a north-facing slope of Jeff Davis Peak, to $-7.76^\circ$C from a protalus rampart in North Fork Baker Creek. The calculated temperature depression estimates have a mean value of $-6.61^\circ$C (Table 2.4). Temperature depression estimates using lapse rates from Greenstein (1983) and Wolfe (1992) range from $-3.88^\circ$C, calculated at a protalus rampart in Teresa Cirque, to $-6.43^\circ$C from solifluction lobes on the north-facing slope of Jeff Davis Peak, with a mean value of $-5.32^\circ$C (Table 2.4). Temperatures calculated using both lapse rates are plotted against one another in Figure 2.14 and have a correlation coefficient of 0.84. The calculated mean values of $-6.61^\circ$C and $-5.32^\circ$C for Full Glacial cooling are within the range of previously calculated full glacial cooling estimates of Broecker and Orr (1958),
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<td>-1.3</td>
<td>-6.55</td>
<td>2.43</td>
<td>-5.42</td>
</tr>
<tr>
<td>MM</td>
<td>Solifluxion Lobes</td>
<td>3392</td>
<td>-0.43</td>
<td>-7.42</td>
<td>3</td>
<td>-4.85</td>
</tr>
<tr>
<td>MMT</td>
<td>Stone Polygons</td>
<td>3358</td>
<td>-0.17</td>
<td>-7.68</td>
<td>3.17</td>
<td>-4.68</td>
</tr>
<tr>
<td>MMT</td>
<td>Stone Circles</td>
<td>3362</td>
<td>-0.2</td>
<td>-7.65</td>
<td>3.15</td>
<td>-4.7</td>
</tr>
<tr>
<td>MMT</td>
<td>Cryoplanation Terrace</td>
<td>3367</td>
<td>-0.24</td>
<td>-7.61</td>
<td>3.12</td>
<td>-4.73</td>
</tr>
<tr>
<td>NFBC</td>
<td>Protalus Rampart</td>
<td>3347</td>
<td>-0.15</td>
<td>-7.76</td>
<td>3.22</td>
<td>-4.63</td>
</tr>
<tr>
<td>NFBC</td>
<td>Protalus Rampart</td>
<td>3369</td>
<td>-0.25</td>
<td>-7.6</td>
<td>3.11</td>
<td>-4.74</td>
</tr>
<tr>
<td>TC</td>
<td>Protalus Ramparts</td>
<td>3198</td>
<td>1.04</td>
<td>-6.81</td>
<td>3.97</td>
<td>-3.88</td>
</tr>
<tr>
<td>WPR</td>
<td>Cryoplanation Terrace</td>
<td>3700</td>
<td>-2.77</td>
<td>-5.08</td>
<td>1.46</td>
<td>-6.39</td>
</tr>
<tr>
<td>WPR</td>
<td>Cryoplanation Terrace</td>
<td>3321</td>
<td>0.11</td>
<td>-7.74</td>
<td>3.35</td>
<td>-4.5</td>
</tr>
</tbody>
</table>

Mean ---> -6.610417 -5.32125


** Lapse Rates (D) are derived from Dohrenwend (1984), (W) are derived from Greenstein (1983) and Wolfe (1992).

Table 2.4: Lapse rates and climate data used to calculate MAAT and temperature depression.
Figure 2.14: Regression of temperature estimates calculated using lapse rates from Dohrenwend (1984) versus temperature estimates calculated using lapse rates from Greenstein (1983) and Wolfe (1992).

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Brackenridge (1978), and Dohrenwend (1984). Temperature estimates here were calculated using a 30-m DTM for the underlying calculations. Temperatures were calculated for each periglacial feature, with 30 meters added and subtracted to the actual elevation to assess the range of error introduced by the digital terrain model. A comparison of the results using all three elevation ranges is provided in Table 2.5.

The spatial distribution of rock glaciers and smaller periglacial features are strongly influenced by potential direct radiation (Funk and Hoelzle, 1992; Hoelzle, 1992; Hoelzle, 1996; and Frauenfelder et al., 2001). Potential direct radiation was calculated for the lower limit of each rock glacier and for the mean elevation of each periglacial feature identified in the Snake Range at the winter and summer solstices (Figs. 2.15 and 2.16). Winter potential direct radiation values range from 0.0 to 8.07 MJm"2d"1 and summer radiation values range from 14.68 to 22.42 MJm"2d"1. Multiple regression analysis of the solar insolation data implies that during the winter, potential direct solar radiation decreases with increasing elevation and that during the summer, potential direct solar radiation increases with increasing elevation (Fig. 2.17). This correlation is best

<table>
<thead>
<tr>
<th>Method</th>
<th>Mean T_D</th>
<th>Mean T_W</th>
<th>ΔT_D +30m</th>
<th>ΔT_D -30m</th>
<th>ΔT_W +30m</th>
<th>ΔT_W -30m</th>
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</thead>
<tbody>
<tr>
<td>Elev</td>
<td>-6.61°C</td>
<td>-5.32°C</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Elev +30</td>
<td>-6.40°C</td>
<td>-5.48°C</td>
<td>+0.26°C</td>
<td>-</td>
<td>-0.19°C</td>
<td>-</td>
</tr>
<tr>
<td>Elev -30</td>
<td>-6.75°C</td>
<td>-5.18°C</td>
<td>-</td>
<td>-0.09°C</td>
<td>-</td>
<td>+0.11°C</td>
</tr>
</tbody>
</table>

Table 2.5: Comparison of possible error introduced by using a 30-meter digital terrain model to estimate Full Glacial cooling. (Elev+30 = original DEM elevation plus 30 meters, Elev-30 = original DEM elevation minus 30 meters, and the ΔT +/- 30m values represent the estimated error in temperature estimates caused by possible errors in the elevation estimates).
Figure 2.15: Periglacial features versus winter radiation values illustrating the relatively low radiation values intercepted by rock glaciers.
Figure 2.16: Periglacial features versus summer radiation values.
Figure 2.17. The relationship between winter and summer radiation versus elevation.

**Summer Radiation vs Elevation Regression Results:** $r^2 = 0.4363; y = 2664.61417 + 26.4648455x$

**Winter Radiation vs Elevation Regression Results:** $r^2 = 0.5596; y = 3281.79935 - 29.9117799x$
Figure 2.18: A surface plot of a multivariate regression between elevation and summer and winter radiation values.
illustrated as a 3D surface plot illustrating the summer and winter solstice radiation values at any given elevation (Fig. 2.18). This relationship most likely reflects changes in aspect and greater topographic interference at higher elevations in the range.

**Equilibrium Line Altitude Reconstruction**

Two episodes of glaciation have been identified and described in the interior Great Basin. These are the Angel Lake and Lamoille advances, which correlate to Tioga and Tahoe advances, respectively, based primarily on geomorphic characteristics and stratigraphic position (Blackwelder, 1931; Blackwelder, 1934; Sharp, 1938; Wayne, 1984; Peigat, 1980; Osbom, 1990; and Osborn and Bevis, 2001). The spatial extents of Angel Lake and Lamoille glaciers in the Snake Range were reconstructed using a combination of field evidence and aerial photography mapping (Porter, 1975). A variety of geomorphic features were used to delineate the extent of glaciation in each valley, including terminal and lateral moraines, changes in valley morphology, trimlines, and truncated spurs. Calculated values for the elevation of Angel Lake glacial ice in the Snake Range have a mean of 3341 masl, and the mean of the lowest terminal extent of Angel Lake glaciers is 2783 masl. The mean of the highest extent of Lamoille glacial ice is 3349 masl, and the mean of the lowest extent of ice is 2478 masl. There is significant uncertainty in the actual spatial extent of some Angel Lake glaciers and in the lower limit of most Lamoille glaciers because many of the glaciers coalesced, making the actual lowermost limit difficult to delineate (Fig. 2.19). The Lehman glacier developed in the most protected and northerly facing cirque, and is assumed to have been more robust than the Stella and Teresa systems. However, it most likely had a smaller accumulation area and, unlike the Teresa and Stella lobes, it was confined to steep valley walls. It is possible that the Teresa and Stella lobes could have been more extensive.
Figure 2.19: A diagram illustrating the geometry of the Lehman Creek glacial lobe. It is difficult to delineate the actual extent and influence of the Lehman, Teresa and Stella lobes that contributed to the overall geometry of the full valley lobe.
because they had a larger accumulation area and breached their respective cirque basins at a higher elevation than did the Lehman lobe. The Lehman Creek glacial system most certainly coalesced into a large valley glacier, however differentiating between the lobes emanating from Stella, Teresa, and Lehman cirques is highly subjective. A summary of the characteristics for each reconstructed Angel Lake and Lamoille glacier in the Snake Range is provided in Table 2.6.

Absolute ages for each glaciation are lacking because of a number of factors: (1) the moraines lack organic material suitable for radiocarbon dating, (2) the coarse resistant Prospect Mountain Quartzite does not weather into material suitable for luminescence dating, and (3) lateral and terminal moraines lack suitable boulders for cosmogenic surface dating. Most Lamoille-age terminal moraines have been eroded, and those that remain have been significantly dissected. Relative ages calculated from weathering parameters were not possible because of the lack of well-developed patina on Prospect Mountain Quartzite, with the exception of Snake Creek that is intruded by the Snake Creek–Williams Canyon pluton.

Three techniques were used to reconstruct former ELAs in the Snake Range: the toe-to-headwall altitude ratio method (THAR), the accumulation area ratio method (AAR), and the balance ratio method (BR). Reconstructed AAR and BR ELAs for glaciers in Lehman and Baker Creek are considered tentative because of sparse geomorphic evidence to accurately delineate accumulation area zones. The mean ELAs calculated using each technique and the statistics that were calculated from first and second order trend interpolations are summarized in Table 2.7.

There is a high correlation between the three methods (THAR, AAR, BR) primarily because each method uses the same 20–22 reconstructed glaciers, so they share similar
<table>
<thead>
<tr>
<th>Location</th>
<th>Aspect</th>
<th>Area (m$^2$)</th>
<th>Length (km)</th>
<th>Area (m$^2$)</th>
<th>Length (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Southern Snake Range</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Johnson Cirque</td>
<td>094</td>
<td>1366516.281</td>
<td>2.690</td>
<td>5658556.754</td>
<td>19619.471</td>
</tr>
<tr>
<td>Shoshone Cirque</td>
<td>064</td>
<td>1511089.477</td>
<td>2.700</td>
<td>5658556.754</td>
<td>19619.471</td>
</tr>
<tr>
<td>Washington Cirque</td>
<td>089</td>
<td>425511.256</td>
<td>1.650</td>
<td>5658556.754</td>
<td>19619.471</td>
</tr>
<tr>
<td>N Fork Baker Creek</td>
<td>049</td>
<td>2452606.111</td>
<td>3.670</td>
<td>5608085.329</td>
<td>20524.802</td>
</tr>
<tr>
<td>Baker Cirque</td>
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<td>1964349.264</td>
<td>3.200</td>
<td>5608085.329</td>
<td>20524.802</td>
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<tr>
<td>Bald Mountain N</td>
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<td>312850.529</td>
<td>0.980</td>
<td>661206.042</td>
<td>5869.889</td>
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<tr>
<td>Bald Mountain E</td>
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<td>827827.928</td>
<td>1.970</td>
<td>2444781.624</td>
<td>10647.009</td>
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<tr>
<td>Bald Mountain W</td>
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<td>553435.760</td>
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<td>755738.038</td>
<td>5451.302</td>
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<tr>
<td>Williams Canyon</td>
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<td>2.610</td>
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<td>N/A</td>
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<td>N/A</td>
<td>N/A</td>
<td>N/A</td>
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<tr>
<td>Pyramid Peak E</td>
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<td>2.870</td>
<td>N/A</td>
<td>N/A</td>
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<tr>
<td>Pyramid Peak W</td>
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<td>283698.126</td>
<td>2.070</td>
<td>N/A</td>
<td>N/A</td>
</tr>
<tr>
<td>Baker Creek Ridge</td>
<td>041</td>
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<td>N/A</td>
<td>N/A</td>
<td>N/A</td>
</tr>
<tr>
<td>Stella Lake</td>
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<td>3621805.651</td>
<td>2.760</td>
<td>5723071.627</td>
<td>17508.458</td>
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<tr>
<td>Teresa Cirque</td>
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<td>3621805.651</td>
<td>2.940</td>
<td>5723071.627</td>
<td>17508.458</td>
</tr>
<tr>
<td>Lehman Cirque</td>
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<td>3621805.651</td>
<td>3.300</td>
<td>5723071.627</td>
<td>17508.458</td>
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<td>Decathlon Canyon</td>
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<td>330438.169</td>
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<td>N/A</td>
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<tr>
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<td>N/A</td>
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<td>N/A</td>
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<td>N/A</td>
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<tr>
<td>Granite Peak</td>
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<td>N/A</td>
<td>N/A</td>
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<tr>
<td><strong>Northern Snake Range</strong></td>
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<td></td>
<td></td>
<td></td>
<td></td>
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<tr>
<td>Big Canyon</td>
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<td>1829843.923</td>
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<td>N/A</td>
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<td>N/A</td>
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<td>Mt Moriah 2</td>
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<td>N/A</td>
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<td>Deadman Creek</td>
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<td>675469.679</td>
<td>3.090</td>
<td>N/A</td>
<td>N/A</td>
</tr>
<tr>
<td><strong>MEAN</strong></td>
<td></td>
<td>1250689.661</td>
<td>2.501818182</td>
<td>4474798.319</td>
<td>15854.69009</td>
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</tbody>
</table>

Table 2.6: Summary of the characteristics of reconstructed Angel Lake and Lamoille glaciers.
<table>
<thead>
<tr>
<th>Method</th>
<th>Number of Cirques</th>
<th>Mean Altitude (m)</th>
<th>Local Polynomial First Order Surface RMSE (m)</th>
<th>Local Polynomial Second Order Surface RMSE (m)</th>
<th>Global Polynomial First Order Surface RMSE (m)</th>
<th>Global Polynomial Second Order Surface RMSE (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>THAR\textsubscript{A}</td>
<td>22</td>
<td>3065.65</td>
<td>112.9</td>
<td>0.0084</td>
<td>159.2</td>
<td>0.9777</td>
</tr>
<tr>
<td>THAR\textsubscript{L}</td>
<td>11</td>
<td>2909.68</td>
<td>101.9</td>
<td>0.2655</td>
<td>80.6</td>
<td>0.9902</td>
</tr>
<tr>
<td>AAR</td>
<td>20</td>
<td>3034.59</td>
<td>125.9</td>
<td>0.7927</td>
<td>108.8</td>
<td>0.0781</td>
</tr>
<tr>
<td>BR</td>
<td>20</td>
<td>2895.98</td>
<td>147.9</td>
<td>0.2655</td>
<td>230.9</td>
<td>0.108</td>
</tr>
</tbody>
</table>

**Comparison of ELA Calculation Methods**

<table>
<thead>
<tr>
<th>Method</th>
<th>n</th>
<th>r</th>
<th>r^2</th>
<th>p</th>
</tr>
</thead>
<tbody>
<tr>
<td>AAR vs THAR</td>
<td>20</td>
<td>0.745</td>
<td>0.555</td>
<td>0.0002</td>
</tr>
<tr>
<td>AAR vs BR</td>
<td>20</td>
<td>0.707</td>
<td>0.5</td>
<td>0.0005</td>
</tr>
<tr>
<td>THAR vs BR</td>
<td>20</td>
<td>0.868</td>
<td>0.754</td>
<td>7E-07</td>
</tr>
</tbody>
</table>

* Global Polynomial Interpolation Parameters (81% Global and 19% Local)

Table 2.7: Summary of mean ELA's and summary statistics calculated using the THAR, AAR and BR methods.
Figure 2.20: Comparison of ELAs calculated using the AAR and THAR methods.
Figure 2.21: Comparison of ELAs calculated using the BR and THAR methods.
Figure 2.22: Comparison of ELAs calculated using the BR and AAR methods.

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area to elevation characteristics (Figs. 2.20, 2.21, 2.22). The only major exception is the Decathlon Canyon glacier, which has the highest calculated ELA in the Snake Range, probably because this canyon faces directly south.

**Paleoclimatic Significance of Estimated Equilibrium Line Altitudes**

Equilibrium line altitudes calculated using different methods for twenty reconstructed Angel Lake glaciers in the Snake Range exhibit a high correlation. The mean ELA from each method was subtracted from the modern ELA of 3673 masl at the Lehman Glacier on Wheeler Peak to calculate ELA depression since the last Full Glacial and the temperature change associated with this depression. Temperature estimates were calculated using lapse rates from Dohrenwend (1984) and Greenstein (1983) and Wolfe (1992). The estimates using Dohrenwend's lapse rate of -0.76°C/km provides the most realistic values and are assumed to most accurately reflect paleo-temperature (Table 2.8).

<table>
<thead>
<tr>
<th>Location</th>
<th>THAR_{AL 0.50}</th>
<th>AAR_{0.65}</th>
<th>BR_{0.65}</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mean</td>
<td>3065.68</td>
<td>3024.53</td>
<td>2909.98</td>
</tr>
<tr>
<td>Modern ELA</td>
<td>3673.00</td>
<td>3673.00</td>
<td>3673.00</td>
</tr>
<tr>
<td>ELA Depression</td>
<td>607.32</td>
<td>648.47</td>
<td>763.02</td>
</tr>
<tr>
<td>$\Delta T_D$</td>
<td>-4.55°C</td>
<td>-4.86°C</td>
<td>-5.72°C</td>
</tr>
<tr>
<td>$\Delta T_W$</td>
<td>-3.04°C</td>
<td>-3.24°C</td>
<td>-3.82°C</td>
</tr>
</tbody>
</table>

| Angel Lake Mean Value $\Delta T_D$ | -5.05°C |
| Angel Lake Mean Value $\Delta T_W$ | -3.36°C |

**Table 2.8:** Mean ELA values, estimated ELA depression since the last Full Glacial, and the calculated temperature change associated with this depression. The mean temperature change using each lapse rate (Dohrenwend, 1984 and Wolfe, 1992) is also provided.
The THAR method indicates an average ELA depression of 607 meters with a temperature change of -4.55°C. The AAR method calculates an average ELA depression of 573 meters with a temperature change of -4.29°C, and the BR method suggests an ELA depression of 763 meters with a calculated temperature change of -5.72°C. These values are consistent with previously calculated temperature depression estimates for the last Full Glacial in the western Great Basin (Table 2.9).

<table>
<thead>
<tr>
<th>Reference</th>
<th>Basis for estimate</th>
<th>Location</th>
<th>Δ Temp °C</th>
</tr>
</thead>
<tbody>
<tr>
<td>Broecker and Orr, 1958</td>
<td>Hydrologic budget</td>
<td>Lake Lahonton, NV</td>
<td>-5.0</td>
</tr>
<tr>
<td>Snyder and Langbein, 1962</td>
<td>Snow line lowering</td>
<td>Lake Spring, NV</td>
<td>-3.6</td>
</tr>
<tr>
<td>Brackenridge, 1978</td>
<td>Snow line lowering</td>
<td>Southwest, USA</td>
<td>-7.0</td>
</tr>
<tr>
<td>Mifflin and Wheat, 1979</td>
<td>Hydrologic budget</td>
<td>Pluvial lakes in NV</td>
<td>-2.8</td>
</tr>
<tr>
<td>Dohrenwend, 1984</td>
<td>Nivation landforms</td>
<td>Western GB</td>
<td>-7.0</td>
</tr>
<tr>
<td>Benson, 1986</td>
<td>Pluvial lakes</td>
<td>Western GB</td>
<td>-7.0</td>
</tr>
<tr>
<td>Benson and Thompson, 1987</td>
<td>Pluvial lakes</td>
<td>Western GB</td>
<td>-7.0 to -10</td>
</tr>
<tr>
<td>Bevis, 1995</td>
<td>ELA estimates</td>
<td>Western GB</td>
<td>-4.0 to -8.0</td>
</tr>
</tbody>
</table>

Mean ⇒ -5.86

Table 2.9: Summary of previously calculated temperature depressions during the last Full Glacial in the western United States.

Conclusions

This study presents new information regarding the location and development of permafrost in the high elevations of the Snake Range, Nevada. It is not surprising that permafrost features are present at such high elevations in a mountain range with the only modern glacier in the interior Great Basin and a very apparent history of late Pleistocene glaciation. Many of the periglacial landforms identified in this study have been described in mountain ranges throughout the southwest (Smith, 1936; Richmond, 1962; Blagbrough and Breed, 1967; Blagbrough and Farkas, 1968; Blagbrough and Breed, 1969; Galloway, 1970; Blagbrough, 1971; Barsch and Updike, 1971; Péwé and Updike,
Neoglacial temperature depression estimates, calculated using modern freezing and thawing indices for relict rock glaciers range from $-0.25^\circ$C to $-1.00^\circ$C. These estimates are anomalously low and are thought to reflect a winter snow cover greater than 50 cm. However, the temperature gradient illustrated by these estimates is identical to the temperature gradient calculated using temperature estimates defined using MAAT and potential direct solar radiation. The temperature estimates calculated using the methodology described by Frauenfelder and Kääb (2000) and Frauenfelder et al. (2001) are $0.35^\circ$ to $0.97^\circ$C lower than those calculated using freezing and thawing indices. The calculated mean of $-1.42^\circ$C for the higher estimates are more consistent with previously published Neoglacial temperature estimates for the Great Basin (Wayne, 1984). However, the MAAT calculated for each rock glacier terminus could not be corrected for the effect of topographic shielding or local orographically-induced microclimates that could significantly depress the MAAT.

Using temperature thresholds for continuous, discontinuous and sporadic permafrost, as defined by Billings and Mooney (1968), the MAAT for relict permafrost features is assumed to be approximately $-7.8^\circ$C. Comparing modern and Full Glacial MAAT in the Snake Range, calculated using regional lapse rates, results in an average depression of the MAAT by approximately $-5.16^\circ$C to $-6.61^\circ$C. These estimates assume that winter precipitation in the Great Basin during the last full glacial was not significantly greater than present. This assumption is supported by previous paleoclimate studies in the southwest (Galloway, 1970; Brackenridge, 1978; Spaulding et al., 1983; and Zielinski
and McCoy, 1987). Not only do these estimates of full glacial MAAT compare favorably with other estimates for this region (Table 2.9), they are also similar to the MAAT depression of $-4.55^\circ C$ to $-5.77^\circ C$ calculated from reconstructed Angel Lake ELAs.

Temperature estimates from both relict permafrost features and ELA depression are consistent with previously published temperature gradients (Bevis, 1995). Bevis (1995) calculated MAAT depressions of $-8.0^\circ C$ in the Ruby Mountains northwest of the Snake Range, $-5.0^\circ C$ in the Deep Creek Mountains, to the northeast of the Snake Range, and $-4.0^\circ C$ in the Tushar Mountains in south-central Utah. Estimates from landforms in the Snake Range are consistent with the southeasterly depression of Full Glacial MAAT estimates described by Bevis (1995). Periglacial deposits in the Snake Range are inactive and appear to be relict features that most likely co-evolved with the onset of alpine glaciation, although it is possible that some features were active during the Holocene. Many of the landforms are similar in size and distribution as those previously described in the Colorado Front Range of Wisconsin age (Pewe, 1983). The presence of these features indicates that an extensive periglacial environment existed in the higher elevations and protected niches of the Snake Range during the last Full Glacial and early deglaciation.

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CHAPTER 3

GEOMORPHIC AND GPR EVIDENCE FOR THE POSSIBLE GENESIS OF A TONGUE SHAPED ROCK GLACIER, SOUTHERN SNAKE RANGE, NEVADA

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Introduction

Rock glaciers have been recognized and studied as discrete geomorphic landforms for over a century. While many questions concerning the conditions for their initial development, evolution, and preservation have been investigated, the debate over a glacial or periglacial origin is still controversial even though there is increasing evidence these landforms evolve in both environments (Krainer and Mostler, 2000; Humlum, 2000; and Ishikawa et al., 2001).

The term rock glacier refers to a tongue or lobate shaped body of angular boulders and fine material that contains either periglacial ice filling subsurface voids, (i.e.– ice cemented), or buried glacial ice, (e.g.– ice cored), exhibiting ridges and furrows and has a steep frontal snout at the angle of repose. Discussion by supporters of the “permafrost” origin can be found in the following references (Wharhaftig and Cox, 1959; Barsh, 1987; 1988; 1992; and 1996; Berthling et al., 2000; Cui and Cheng, 1988; Gorbunov, 1983; Haeberli and Schmid, 1988; Haeberli et al., 1988; 1992; 1998; Isaksen et al., 2000; Ishikawa, et al., 2001; Jakob, 1992; 1994; King et al., 1987; Vonder Mühll and Haeberli, 1990; Vonder Mühll and Schmidt, 1993), while rock glaciers attributed to a “glacial origin” are discussed by Benedict et al. (1986), Calkin et al. (1987), Clark et al. (1994 and 1996), Ellis and Calkin (1979), Embleton and King (1975), Evans (1993), Eyles (1978), Gardner (1978), Haeberli (1989), Humlum (1982; 1988a; 1988b; and 1996), Ishikawa, et al. (2001), Johnson and Lacasse (1988), Krainer and Mostler (2000), Lliboutry (1953 and 1990), Martin and Whalley (1987), Outcalt and Benedict (1965), Potter (1972), Whalley (1974 and 1992), Whalley and Martin (1992), Whalley et al. (1994 and 1995) and White (1975). However, proponents of both origins generally agree that active rock glaciers develop in alpine areas characterized by high rates of talus
production and local permafrost conditions (Corte, 1987). Thus, alpine environments with high topographic relief are ideal locations because they provide the necessary talus and local climatic conditions for the development of a rock glacier. Rock glaciers typically exhibit two stratigraphically and texturally distinct layers: an upper, active layer (1 – 5 m thick) composed of blocky debris, and a lower layer (20 – 30 m thick) composed of ice (either massive or interstitial) and talus derived debris. However, there is still uncertainty regarding the specific mechanics responsible for the initiation and genesis of rock glaciers (Humlum, 1998, 2000).

Considerable research has been conducted to address the long-standing debate concerning whether rock glaciers should be classified as glacial or periglacial geomorphic landforms. A first-order approach involves investigating the internal composition of rock glaciers to identify whether they contain discontinuous lenses of interstitial ice or a remnant core of glacial ice. Previous research on the internal ice content and structure of rock glaciers has involved direct sampling of drill cores, trenching or seismic reflection requiring explosives (White, 1971; Barsch et al., 1979; Haeberli et al., 1988; Cui and Zhu, 1989; Vonder Mühll and Holub, 1992; Barsch, 1996; Haeberli and Vonder Mühll, 1996; Elconin and La Chapelle, 1997; Fisch et al., 1997; Konrad et al., 1999; Vonder Mühll et al., 2000; Musil et al., 2002; and Żurawek, 2002). However, the study area is located in Great Basin National Park (GBNP), therefore any attempt to investigate the internal characteristics of the rock glacier required a non-intrusive and non-destructive technique. Previous research has illustrated the merits of using ground-penetrating radar (GPR) to investigate the internal morphology of rock glaciers and talus rich permafrost (Berthling et al., 2000; and Isaksen et al., 2000; Degenhardt et al., 2000; Degenhardt et al., 2001a; 2001b; Hinkel et al., 2001; Volkel et
al., 2001; Sass and Wollny, 2001; and Degenhardt et al., 2003). Therefore, we felt confident using GPR as an alternative for the non-destructive evaluation of subsurface structures in the Lehman rock glacier.

The goals of this study were to: (1) investigate the geomorphology of the Lehman rock glacier, (2) improve our understanding of the processes controlling the development of the Lehman rock glacier by investigating the internal structure using GPR, and (3) use geomorphic and GPR data to evaluate our working hypothesis regarding the genesis of the Lehman rock glacier. This study is not an attempt to resolve the debate over a glacial versus periglacial origin for rock glaciers. Rather, we present results from a study of the Lehman rock glacier in the southern Snake Range, Nevada and suggest an alternative “recessional genesis” for some tongue-shaped rock glaciers in alpine regions.

Study Area

The southern Snake Range is located in east-central Nevada in the Basin and Range physiographic province (Fig. 3.1). The Snake Range is an uplifted horst, bounded to the east by Snake Valley and to the west by Spring Valley. Alpine environments throughout the Basin and Range produce drastically different microclimates from the surrounding lower elevation valleys. Meteorological measurements have been recorded at the Lehman Cave weather station, located at an elevation of 2073 meters above sea level (masl), in Great Basin National Park since 1961. The modern mean annual air temperature (MAAT) recorded at the Lehman Cave weather station is 9.59°C; the coldest month is January (−2.69°C) and the warmest month is July (22.75°C). Mean annual precipitation at the Lehman Cave climate station is approximately 2.40 cm; the greatest amount of precipitation falls in May (3.23 cm) and the least falls in July (1.63 cm).
Figure 3.1: Location map of the Southern Snake Range with a shaded relief map showing the spatial relationship of the Lehman rock glacier with the surrounding topography.
Average mean annual temperatures for the entire park can range from $-8^\circ C$ to $30^\circ C$, and average precipitation from the valleys to the high peaks region can vary from less than 12 cm to greater than 50 cm (Houghton et al., 1969).

Glaciated portions of the southern Snake Range are underlain by early Paleozoic Prospect Mountain quartzite, Pole Canyon limestone, and Pioche shale intruded by the Snake Creek–Williams Canyon pluton that is Jurassic in age (Drewes, 1958; Whitebread, 1969; Lee et al., 1970; Lee and Van Loenen, 1971; Lee et al., 1981; Lee and Christiansen, 1983 a, b; Lee et al., 1986; Miller, 1995 and Miller, 1999). Glacial features of the Snake Range were first recognized and described by early explorers (Gilbert, 1875; Simpson; 1876; Russell, 1884; Blackwelder, 1931; and Flint, 1947), and subsequent authors have continued to substantiate earlier reports and discuss the glacial history and the Lehman rock glacier (Heald, 1956; Kramer, 1962; Curry, 1969; Piegat, 1980; Waite, 1974; Osborn, 1988; Osborn and Bevis, 2002). However, little research has focused on the geomorphology, genesis, and significance of the Lehman rock glacier. Preliminary reports have described the surface morphology and spatial extent of the Lehman rock glacier (Kramer, 1962; Osborn, 1988; Osborn, 1990; Van Hoesen and Orndorff, 2001; and Osborn and Bevis, 2002).

However, the internal structure of the rock glacier has never before been investigated, and therefore it was uncertain whether it contained interstitial or massive ice. Without understanding the internal structure and ice distribution within rock glaciers it is impossible to thoroughly test how they develop and whether they are inactive or relict landforms.
Methods

We used aerial photographs and field measurements to identify and map the geomorphology of the Lehman rock glacier. Color aerial photographs (1:24,000 and 1:12,000 scale) were obtained from the United States Department of Agriculture, and black and white aerial photographs were obtained from the United States Geological Survey (1:48,000). We identified and mapped surface features on aerial photos using a geographic information system (GIS). During the summers of 2001 and 2002, these features were later compared with field observations.

We describe and classify the rock glacier using the taxonomy of Barsh (1987) and Corte (1987). The morphometric parameters of the rock glacier were calculated in a GIS following the guidelines described by Barsh (1996). We also conducted a fabric analysis (Wahrhaftig and Cox, 1959; Eskenasy, 1978; Giardino and Vitek, 1985; Perez, 1989; Nicholas, 1994; and Thompson, 1999) of talus blocks (>0.5 m) located on three prominent lobes of the Lehman rock glacier.

We used a Noggin Plus GPR system manufactured by Sensors and Software with 250 and 500 MHz antennas to collect GPR profiles from individual lobes of the Lehman rock glacier over the period 19–23 August 2002. Conducting the survey at this time of the year provided the best opportunity to use GPR because of increased access to the rock glacier and low snow cover. The Noggin Plus instrument consists of a digital video logger (DVL), two antennas, and a 12-volt battery. Selected GPR traces were processed using WinEKKO v.1.0 Pro and Transform v.3.4 from Sensors and Software (Mississauga, ON, Canada).

The GPR antennas were hand carried ~ 0.25 – 0.5 m above the surface because large blocks of quartzite talus cover the surface of the rock glacier lobes (Fig. 3.2).
Figure 3.2: Typical surface of the Lehman Cirque rock glacier, composed of large blocks of quartzite.
The GPR unit was carried over a total of four transects; of the transects, three were perpendicular to the flow of each lobe and the fourth was parallel to the long axis of the upper lobe of the rock glacier (Fig. 3.3). Using the 500 MHz antenna we collected GPR data for 4 depths: 38 m, 20 m, 10 m, and 5 m, using the 250 MHz antenna we collected GPR data for 2 depths: 78 m and 45 m. Survey tape was used to mark 2.5 m intervals along each transect. As the antenna passed over each 2.5-meter point, a fiduciary marker was inserted into the radar profile. We were unable to obtain GPS coordinates for topographic correction in the field because of steep topography in the upper portion of the cirque. We attempted to extract GPS coordinates from a 30-meter USGS digital elevation model (DEM) for the middle and upper lobes; however the resolution of the DEM provided inadequate data for topographic correction. Table 3.1 summarizes the location, antenna frequency, length, selected depth of penetration and the orientation of each GPR profile.

Results

Geomorphology

The Lehman Cirque rock glacier is approximately 900 m long, located in a steep north-to-northeast facing cirque composed of Prospect Mountain Quartzite. The rock glacier is in direct connection with the source area and it has an estimated rooting zone elevation of 3667 masl (Fig. 3.3). The rock glacier is a complex feature containing three distinct lobes mantled by large (0.5 to 3.0 m) quartzite boulders and composed of poorly-sorted, fine-grained material. The surface of the rock glacier exhibits well-developed relief (2.0 – 3.0 m) in the form of prominent furrows and ridges that are oriented perpendicular to down-valley motion (Fig. 3.4). Most of the ridges have
Figure 3.3: United States Department of Agriculture aerial photograph (1:12,000) of the Lehman rock glacier showing the location of GPR transects and modern and former rock glacier initiation line altitudes (RILA).
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rounded crests, and the furrows are asymmetrical \(v\)-shaped features. Both the furrows and ridges are composed of cobble-to-boulder-sized debris. The lateral margins of the rock glacier are steep, and the upper surface of each lobe is generally convex with a subtle southwest-to-northeast gradient.

Absolute ages are not available for the Lehman Cirque rock glacier for a number of reasons: (1) organic material suitable for radiocarbon dating has not yet been recovered, (2) the frost-riven nature of the quartzite boulders renders them unsuitable for cosmogenic dating, and (3) low weathering rates of the quartzite debris does not provide suitable material for luminescence dating. However, previous work suggests an age of 1200AD or younger for the Lehman Cirque rock glacier, based on the presence of Mono Crater ash, and the upper lobe is thought to have formed during the Little Ice Age (Osborn, 1990 and Osborn and Bevis, 2002).

The upper lobe is a convex landform approximately 434 m long and 206 m wide. It exhibits well-developed arcuate furrows and ridges ranging in height from 1.5 – 3.0 meters. The frontal and lateral margins of the lobe are composed of light gray sediment and talus. They are steep (34 – 36\(^\circ\)), and capped by a 0.5 m thick sequence of dark brown talus. The frontal margin of the lobe exhibits poorly developed boulder aprons and sparse vegetation is found growing on the surface. At the base of the terminus of this lobe there is a well-developed, 3-4-meter-deep, arcuate trench that is oriented perpendicular to the flow of the rock glacier (Fig. 3.4). Cobbles exposed in the trench between the upper and middle lobe are dipping to the northeast 25 – 27\(^\circ\).

The middle lobe is a moderately convex landform, approximately 265 m long and 178 m wide with moderately developed arcuate furrows and ridges that range in height from 1
Figure 4.4: Field photos of the Lehman Cirque rock glacier illustrating: (a) the morphology of a deep arcuate furrow at the terminus of the middle lobe, and (b) the location of other arcuate furrows and well-developed furrows and ridges on the upper lobe.
- 2 m. The frontal margin of this lobe is also composed of fine-grained sediment and talus overlain by a darker sequence of blocky quartzite that is approximately 1 m thick. The frontal slopes are steep (32° - 36°), vegetation is slightly more abundant than on the upper lobe, and poorly developed boulder aprons are present at the base of the steep frontal snout. The lateral margins of the lobe are difficult to distinguish because of the presence of abundant talus and the subsequent coalescence of talus and rock glacier margins. At the base of the frontal snout of this lobe there is also a well-developed, 2-3-meter-deep, arcuate depression oriented perpendicular to the flow of the rock glacier. Cobbles exposed in the trench between the middle and lower lobe, are dipping to the northeast 20° - 23°.

The lower lobe is a weakly convex landform. It is approximately 195 m long, 100 m wide, and lacks well-developed surface furrows and ridges. The surface of the lower lobe exhibits swell-and-swale topography, but it is poorly preserved. The frontal margin is composed of fine-grained sediment overlain by weathered talus. The frontal slopes are steep (30° - 33°), well-developed talus aprons are present at the base of this lobe, and the frontal snout is moderately vegetated.

The results of a statistical analysis of fabric data collected on talus blocks covering the surface of the three rock glacier lobes are compared and summarized in Table 3.2. The orientation and vector mean for talus blocks on each lobe are plotted on rose diagrams shown in Figure 3.5. The vector mean ranges from 31° to 358°, the vector resultant length (\( R \)) ranges from 0.92 to 0.97, and the von Mises distribution (\( \kappa \)) ranges from 6.39 to 18.99. The vector resultant length is a measure of dispersion and is analogous to variance, except that values closer to 1.0 indicate that observations are tightly bunched together (Mardia and Jupp, 1999; Fisher, 1993; and Davis, 1986). The
Table 3.2: Statistical variation between the geomorphically distinct Lehman rock glacier lobes.

<table>
<thead>
<tr>
<th>Location</th>
<th>n</th>
<th>R</th>
<th>Mean Vector (μ)</th>
<th>von Mises parameter (κ)</th>
<th>Circular Variance</th>
<th>Circular Standard Deviation</th>
<th>Standard Error of Mean</th>
<th>95% Confidence Interval</th>
<th>F - Test (p)</th>
<th>t-Test (p)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Upper Lobe</td>
<td>83</td>
<td>0.95</td>
<td>47.42°</td>
<td>9.63</td>
<td>0.05</td>
<td>18.99°</td>
<td>2.08°</td>
<td>43.34° ± 51.51°</td>
<td>&lt;0.05</td>
<td>&lt;0.05</td>
</tr>
<tr>
<td>Middle Lobe</td>
<td>72</td>
<td>0.92</td>
<td>358.02°</td>
<td>6.39</td>
<td>0.03</td>
<td>23.69°</td>
<td>2.79°</td>
<td>352.56° ± 3.49°</td>
<td>&lt;0.05</td>
<td>&lt;0.05</td>
</tr>
<tr>
<td>Lower Lobe</td>
<td>75</td>
<td>0.97</td>
<td>31.37°</td>
<td>18.14</td>
<td>0.08</td>
<td>13.65°</td>
<td>1.58°</td>
<td>28.28° ± 34.46°</td>
<td>&lt;0.05</td>
<td>&lt;0.05</td>
</tr>
</tbody>
</table>

F-Test null hypothesis is that the variances are equal

$t$-Test null hypothesis is that mean values are equal
Figure 3.5: Rose diagrams illustrating the orientation of talus block on the upper, middle and lower lobe of the Lehman Cirque rock glacier. Frequency bins were calculated at 10° intervals for the radius of each wedge. The mean vector and 99% confidence intervals are also provided. The statistics for each lobe are summarized in Table 3.
von Mises distribution is equivalent to a normal distribution test and measures both a mean direction ($\theta$) and a concentration parameter ($\kappa$). As $\kappa$ increases, the probability of observing a directional measurement very close to $\theta$ increases (Mardia and Jupp, 1999; Fisher, 1993; and Davis, 1986). Chi-Squared and F-Tests were calculated between each of the lobes and have a probability level less than 0.05.

**Ground Penetrating Radar**

We utilize caution when describing and interpreting the GPR data gathered in this study. We were unable to topographically correct the profiles, calculate a common midpoint profile, or identify parabolic reflectors in any of the profiles. Without a common midpoint profile we were unable to estimate the velocity of the medium or perform semblance analysis to differentiate between air and ice in the upper section of each profile. This is problematic because it prevents us from determining whether the velocity of the reflectors represented the combination of the air wave reflection and evanescent and direct wave, or just the air wave velocity (Degenhardt and Giardino, 2003 and Degenhardt et al., 2003). We assumed the velocity of the medium to be 0.11 m ns$^{-1}$ based on published values for temperate rock glaciers (Degenhardt et al., 2003). This value was used for the depth conversion of each profile. We calculated the theoretical vertical resolution for the 500 MHz and the 250 MHz antennas using the equation $\lambda/4 = (V/F)/4$ (Reynolds, 1997), where $\lambda$ represents the wavelength in meters, $V$ represents the radio wave velocity of the medium (m ns$^{-1}$), and $F$ is the center frequency of the antenna (MHz). The vertical resolution calculated for the 500 MHz antenna is 0.09 m, and the 250 MHz antenna has a theoretical vertical resolution of 0.04 m. Therefore, each profile collected using the 250 and 500 MHz antennas provides exceptional resolution for thin layers and smaller clasts, but lack the resolution to
identify medium to large sized clasts or layers. The wavelengths produced by higher frequency antennas are too short to allow us to differentiate between the small, medium, and large clasts. This study utilized higher frequency antennas than used in previous studies; typical antenna frequencies range from 25 – 100 MHz.

Unfortunately, the Noggin Plus system does not yet have 25, 50, or 100 MHz antennas, which could have provided data on deeper structures in the rock glacier and provided profiles with better resolution of the larger debris fraction. Therefore, our descriptions and interpretations of the GPR data are limited to the thinnest layers of the upper 2 – 15 meters of each lobe. Interpreting high-frequency radar data is often problematic because the observed reflectors may represent primary and/or secondary events, background noise, and attenuation caused by the presence of water or air.

Previous research has discussed the attenuation of high frequency signals caused by high moisture content, sporadic ice content and a thick active layer (Lawson et al., 1998; Berthling et al., 2000; Hinkel et al., 2001; Sass and Wollny, 2001; and Degenhardt and Giardino, 2003), however without the aid of common midpoint and semblance analysis, we were unable to estimate average velocities for different materials represented by the reflectors.

GPR profiles from each lobe reveal a similar network of parallel subsurface reflectors (Fig. 3.6). Each lobe contains three identifiable and distinct layers. The uppermost layer of each GPR depth profile represents the air layer between the Noggin Plus unit and the surface of rock glacier. The second layer, ranging in thickness from 1 – 3 m, is interpreted to represent the active layer, or paleo–active layer, of each lobe and contains reflectors of moderate amplitude that are horizontally discontinuous. The third layer extends from the lower boundary of the active layer to depths ranging from ~10 –
Figure 3.6: GPR profiles collected from the lower, middle and upper lobe. A, B and C were collected perpendicular to the flow of the rock glacier and D was collected parallel to flow. The y axis is measured in nanoseconds and the x axis is measured in meters.
15 m and contains more sporadic, weak reflectors with lower amplitudes and more lateral discontinuity.

**Lower Lobe of the LRG**

The upper 3.0 ns of the lower lobe profile represents the layer of air between the Noggin and the surface of the rock glacier. Below the air layers is an approximately 13.0 ns thick layer containing laterally continuous reflectors that are primarily horizontal and exhibit an ordered layering appearance. These reflectors exhibit the greatest amplitude and contrast of any reflectors identified in the profile (Fig. 3.7). The third layer observed in the radar profile is approximately 13 ns thick and contains weakly dipping (up to ~12°) reflectors that are less laterally continuous. Dipping reflectors are more common near the beginning and end of the profile. Weak discontinuous reflectors are observed down to a depth of 75 ns.

**Middle Lobe of the LRG**

The upper 4.0 ns of the middle lobe profile represents the layer of air between the Noggin and the rock glacier. Below the air layer is an interval that is approximately 4.7 ns thick that contains reflectors that are horizontally and laterally continuous. These reflectors exhibit the greatest amplitude and contrast of any of the reflectors observed in this profile, similar to the lower lobe. The second layer in this profile is approximately 18.0 ns thick, consisting of moderately-to-steeply-dipping (up to ~45°) reflectors that are moderately angular. These reflectors are more discontinuous and truncated than those observed in the lower lobe. Weak and discontinuous reflectors that can be identified to a depth of 35.0 ns characterize the third layer in this profile.

Some reflectors in this profile appear to shallowly dip towards one another, overlain by a region of high signal attenuation that lacks identifiable reflectors (Fig. 3.8). At least
Figure 3.7: Representative portion of the radar profile collected from the lower lobe of the LRG. The bright red and blue horizontal layers represent the air layer between the Noggin and the ground surface. The layer below the air layer containing the reflectors that exhibit the greatest contrast, is thought to represent the paleo-active layer. Strong reflectors are observed down to a depth of ~40 ns and weak reflectors are present down to a depth of ~75 ns. This is only a small segment of the entire GPR profile collected for the lower lobe.
four such structures were identified in this lobe and are randomly distributed throughout the profile.

**Upper Lobe of the LRG**

The upper 4.5 ns of the upper lobe profile represents the air layer between the Noggin unit and the surface of the rock glacier. Below the air layer is an interval that is approximately 4.0 ns thick, consisting predominantly of laterally continuous and horizontal reflectors. These reflectors are more angular in appearance than those identified in the middle and lower lobe. The second layer in this profile is approximately 10 ns thick and contains moderately-to-steeply-dipping (up to ~47°) reflectors that are often chaotic in appearance. They are more chaotic near the edge of the profiles and frequently terminate on other reflectors, creating a random pattern of onlap–offlap structures. Weak, discontinuous reflectors that penetrate to a depth of 30.0 ns characterize the third layer of this profile.

At least two structures in this profile exhibit reflectors with a geometry that is similar to those identified in the middle lobe, with reflectors dipping towards one another overlain by a sequence of strongly attenuated reflectors (Fig. 3.9).

**Discussion**

Previous research has described the movement of rock glaciers in alpine and continental settings (Haeberli, 1985; Barsch, 1996; Whalley and Martin, 1992; and Whalley and Azizi, 1994) and possible models for their genesis (Potter, 1972; Wayne, 1981; Giardino, 1983; Johnson, 1984; Vick, 1987; Haeberli et al., 1998; Whalley and Palmer, 1998; Burger et al., 1999; Konrad et al., 1999; Humlum, 2000; and Isaksen, 2000). Most models address the genesis of rock glaciers from the perspective of a
Figure 3.8: Representative portion of the radar profile collected from the middle lobe of the LRG. The bright red and blue horizontal layers represent the air layer between the Noggin and the ground surface. The layer below the air layer containing the reflectors that exhibit the greatest contrast, is thought to represent the paleo-active layer. Strong reflectors are observed down to a depth of ~35ns and weak reflectors are present down to a depth of ~50ns. The structures identified with arrows are thought to represent thaw pits associated with attenuated reflectors (AR). This is only a small segment of the GPR profile collected from the middle lobe.
Figure 3.9: Representative portion of the radar profile collected from the upper lobe of the LRG. The bright red and blue horizontal layers represent the air layer between the Noggin and the ground surface. The layer below the air layer containing the reflectors that exhibit the greatest contrast, is thought to represent the paleo-active layer. Strong reflectors are observed down to a depth of ~45ns and weak reflectors are present down to a depth of ~70ns. The structures identified with arrows are thought to represent thaw pits associated with attenuated reflectors (AR). This is only a small segment of the GPR profile collected from the upper lobe.
glacial origin. With the exception of Giardino (1983), Johnson (1984), Haeberli et al. (1998) and Isaksen (2000) who propose models of a periglacial origin. Models for the development of ice-cored rock glaciers begin with the accumulation of rockfall debris accumulating on low-angle slopes and at the base of glaciers, eventually developing into a mantle of surface debris that insulates massive ice or snowbanks and initiates the development of a rock glacier (Fig. 3.10). Each successive lobe of the rock glacier is assumed to develop on the underlying debris mantle or existing lobe. This creates a stratification of individual rock glacier lobes with different ages (Fig. 3.11).

Prior workers suggested that the Lehman Cirque rock glacier is composed of only two distinct segments, an upper segment superimposed on a lower and more degraded segment (Heald, 1956; Osborn, 1990; and Osborn and Bevis, 2002). Osborn (1990) interpreted the Lehman Cirque rock glacier to be an ice-cored rock glacier, based on the presence of glacial ice in the rooting zone that grades into the upper lobe of the rock glacier. While it is possible that a transition zone exists at the rooting zone between a debris-covered glacier and the upper lobe of the Lehman Cirque rock glacier, the presence of a debris-covered glacier does not imply the rock glacier is ice cored (Barsch and King, 1989; Jakob, 1992; and Barsch, 1996). Osborn (1990) also describes a steep walled pit on the surface of the rock glacier that retains water during late summer and which bears resemblance to the arcuate pits we describe at the terminus of each lobe. He interprets this as a thaw pit, which he cites as further evidence to support the interpretation of an ice-cored genesis. However, the presence of three arcuate pits with such similar geometry and clasts dipping downvalley does not support the interpretation that these are thaw pits. Clasts exposed in thaw pits should dip towards the center of the pit, not in the direction of downslope movement. Rather, we interpret these pits as
Figure 3.10: Previously published models for the development of the Galena Creek rock glacier, Wyoming and the Marinet rock glacier, France. A: Model published by Konrad et al. (1999). (1) debris begins accumulating (2) Ice overrides existing talus while new talus accumulates on top of the ice, (3) glacier moves downslope transporting the debris layer, and (4) talus migrates over terminus while ice overrides the debris. B: Model proposed by Whalley and Palmer (1998). (1) Advance of the glacier covered by a thin layer of debris, (2) input of talus provides thick debris cover that the glacier carries out of the cirque, (3) the glacier slows down in response to the rock glacier slowing down caused by decreased gradient and decreased ice content, and (4) debris at the terminal snout thickens and provides sufficient insulation to protect the lower ice core.

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Figure 3.11: Schematic diagram illustrating an idealized development of the Lehman rock glacier under the assumption that it developed from a glacial origin using the model proposed by Konrad et al. (1998). (a) The lower lobe of the rock glacier develops by overriding previously deposited debris created an ice cored lobe and (b) the middle lobe of the rock glacier develops in a similar fashion by overriding the upper surface of the previously existing lower lobe. The development of the upper lobe would follow a similar genesis as described for the middle and lower lobe. Applying this model to the Lehman rock glacier is inconsistent with the geomorphic evidence. This "layer cake" stratigraphy is not observed in the field and does not account for the preferred orientation of each lobe or the arcuate pits described at the terminus of the upper and middle lobe.
representing the lowermost terminus of each progressively older lobe. We suggest that individual lobes of the Lehman Cirque rock glacier formed by stacking up behind one another in response to a retreating Lehman glacier during the Neoglacial, creating the appearance of a cohesive tongue-shaped rock glacier (Fig. 3.12).

The limiting factor for the continued development and propagation of each lobe was almost certainly the ice aggradation rate and not a debris supply because the rooting zone is located below a steep headwall and the Lehman Cirque rock glacier is surrounded by steep, well-jointed ridges (Osborn, 1990). Both the steep headwall and ridges surrounding the Lehman Cirque rock glacier provide a more than adequate debris supply.

Geomorphic Evidence

The three distinct lobes of the Lehman Cirque rock glacier are visible both on aerial photos and in the field. The upper lobe is generally accepted as the youngest segment of the rock glacier, and the lower lobe is interpreted to be the oldest portion, based on stratigraphy, degree of preservation, lichenometry studies, extent of surface relief, and tephra chronology (Osborn, 1990 and Osborn and Bevis, 2002). The presence of three distinct lobes suggests instability in the cirque glacier; if the glacier remained stable throughout the late Pleistocene and Neoglacial, then the upper and middle lobes would not have developed. The glacier must have retreated to provide room for each of the lobes to develop (Fig. 3.12). We envision that a small lobate rock glacier developed in Lehman Cirque as the glacier retreated, similar to that seen in Teresa Cirque (Fig. 3.13). A slightly warmer climate would have increased the influx of debris into Lehman Cirque and provided the necessary material to develop the middle and upper lobes during subsequent cooler periods of the Neoglacial. Thus, alternating warm and cool periods
Figure 3.12: Schematic diagram illustrating the hypothesized development of the Lehman rock glacier in response to the receding Lehman glacier (LG). (A) Initial development of the Lehman Glacier, (B) LG recedes and the lower lobe of the LRG develops (I), (C) LG continues to recede and the middle lobe of the LRG develops (II), and (D) the LG recedes to close to its modern location and the upper lobe of the LRG develops. Under modern conditions it appears that the upper lobe is inactive, although it displays evidence of recent activity.
throughout the Neoglacial would have created ideal conditions for the development of individual rock glacier lobes.

Fabric analysis of talus from each of the lobes indicates that all three lobes have distinct longitudinal orientations. Pairwise Chi Squared tests suggest the sample populations are significantly different (at the 0.05 level) and pairwise F-tests (also at the 0.05 level) indicate that this difference is derived from the difference between the vector means of each lobe. Rayleigh uniformity plots provide further evidence supporting our hypothesis that the lobes have a preferred orientation (Fig. 3.14). These plots illustrate that the data from each lobe exhibit a departure from a uniform distribution and have a statistically distinct preferred orientation.

One possible interpretation of these data is that each lobe began to develop and flowed downslope from different locations in the cirque in response to changing local topography as a function of headwall retreat or glacier geometry (Degenhardt et al., 2003). However, if we accept the hypothesis that the Lehman Cirque rock glacier is composed of three distinct lobes and is derived from a glacial origin, this implies that at least three separate bodies of ice developed and were buried in Lehman Cirque, subsequently evolving into the modern landform. Other than the development and rapid burial of three bodies of ice, it is unclear how an ice–cored rock glacier could develop multiple lobes stacked behind one another with different mean orientations. The inference that three separate bodies of ice developed over a 1,200-year (Osborn, 1990) period is unlikely. We suggest that the presence of Mono Crater ash in the lower lobe does not necessarily indicate that this lobe formed no earlier than 1200 AD. It is quite plausible, and more likely that Mono Crater ash fell on the surface of all three lobes of the rock glacier. This ash could very easily infiltrate the porous, blocky talus that forms
Figure 3.13: Small lobate rock glacier located north of Lehman rock glacier in Teresa Cirque. We suggest that the formation of multiple lobate rock glaciers, similar to this feature, could create the appearance of a tongue-shaped rock glacier.
Figure 3.14: Rayleigh uniformity plots illustrating a departure from a uniform distribution. The null hypothesis is that the data are uniform and therefore should plot along the reference line at a 45° angle. Any departure from this line indicates a departure from uniformity and therefore rejects the null hypothesis that the data do not have a preferred orientation.
the upper layer of the lobes, thereby becoming incorporated into the matrix of the lower lobe. This suggests that the lower lobe was still active 1200 years BP, but it is likely that it formed much earlier than this. Its maximum age is constrained by the presence of 5,000-year-old bristlecone pines on a moraine downvalley from the rock glacier (Osborn, 1990). This provides a time frame of 1200 years BP or younger for the upper segment and 5000 years BP or younger for the lower segment, with evidence for downslope activity occurring in the lower lobe approximately 1200 years BP.

We suggest that the variability in lobe orientation reflects a change in the geometry of the Lehman glacier, induced by episodic degradation and possible re-advance in response to a fluctuating Neoglacial climate (Clark and Gillespie, 1997; Dean, 1997; Davis, 1998; Polyak et al., 2001; and Benson et al., 2003). The Lehman Glacier must have retreated multiple times to provide space and sufficient debris for each lobe to develop upslope from the previously developed lobe. We interpret the arcuate features located at the terminus of the upper and middle lobes as evidence that each younger lobe has not overridden the lower, older lobes. Talus and debris from each successively younger lobe is exposed in these arcuate trenches, also suggesting that each older lobe did not develop on the upper surface of a younger lobe. In plan view and in the field, this creates the appearance of a single, cohesive, tongue-shaped rock glacier, when in fact it is composed of multiple tongue or lobate rock glaciers (Fig. 3.15).

By accepting our hypothesis that three stratigraphically distinct rock glacier lobes developed in Lehman Cirque, we also have to accept that three rock glacier initiation line altitudes (RILA) existed throughout the development of the existing landform (Humlum, 1998 and Humlum, 2000). Humlum (1988) defines the RILA as the altitude where rock glaciers creep out from the overlying slope, identified by a break in slope or the upper
Figure 3.15: Possible evolution and genesis of the Lehman rock glacier. (a) Development of small tongue or lobate rock glacier, (b) map view of expected rock glacier morphology, (c) glacier retreats creating room for talus and disrupting the RILA, (d) map view of expected rock glacier and talus morphology, (e) glacier re-advances and initiates the development of a rock glacier in downslope talus and (f) map view of "stacked" small tongue or lobate rock glacier exhibiting the appearance of a single tongue-shaped rock glacier.
limit of debris-covered rock glaciers. The modern RILA can be identified on Figure 3.3 as the transition from the lower portion of the Lehman glacier and the rooting zone of the Lehman Cirque rock glacier. The inferred former RILA for each lobe is also shown in Figure 3.3. A similar environment with a more extensive Lehman Glacier is envisioned for the middle and lower lobes, and the genesis of three well-developed lobes suggests that the development of interstitial segregated ice lenses must have been sufficient to induce downslope movement (Wayne, 1981; Whalley and Martin, 1992; Whalley and Azizi, 1994; and Whalley and Palmer, 1998). However, because of the lack of absolute ages, the temporal relationships between each lobe and the activity of the Lehman Glacier are unclear.

It is possible that the lower lobe represents the Great Basin equivalent of a 4000 BP Neoglacial advance proposed by Davis (1998). Konrad et al. (1999) tentatively correlated the development of the Galena rock glacier in the Absaroka Mountains of Wyoming to this 4000 BP advance. This would suggest that the Lehman Glacier advanced to an elevation of 3333 masl, the current elevation of the modern head of the lower lobe of the Lehman Cirque rock glacier. The middle and upper lobes of the Lehman Cirque rock glacier are certainly younger than the lower lobe, but it is impossible to place absolute constraints on the timing of their development. Osborn (1990) proposed a Little Ice Age origin for the upper lobe. We concur that this is likely, based on the proximity to the modern Lehman glacier and geomorphic evidence suggesting it has recently been active.

Evaluation of GPR Data

Representative radar profiles from each lobe are shown in Figure 3.8. We present profiles that were collected using the 500 MHz antenna set to a depth of 5 m; because we
were unable to image the deepest reflectors, the combination of the highest frequency and shallowest penetration provide the greatest detail. Both the horizontal and vertical axes are expressed in meters. The maximum depth to which we record visible reflectors is 16 m, 15 m and 14 m for the upper, middle and lower lobes respectively, based on an assumed velocity of 0.11 m/ns. The upper (2 – 3 ns) horizontal layers at the top of each profile represent the air layer between the GPR unit and the surface of the rock glacier. Some of the inconsistency in depth of penetration between the radar profiles may be attributed to the speed at which they were collected. Although every attempt was made to maintain a constant speed when carrying the GPR over the surface of the rock glacier, this was impossible due to the differences in terrain between the transects. Hinkel et al. (2001) attributed inconsistencies between the amounts of subsurface penetration to the speed at which the survey was collected. However, the lack of reflectors below variable depths between different lobes using the same frequency antenna suggests inconsistent attenuation along the lower boundary of each GPR profile probably caused by the presence of water.

Radar profiles collected from the lower lobe exhibit more subdued and generally more continuous and horizontally-oriented reflectors than those observed in the middle and upper lobes. However, reflectors are less horizontal near the lateral margins of the rock glacier. This indicates more shearing at the margins caused by increased friction as predicted by model of glacier and rock glacier flow dynamics (Whalley and Martin, 1992 and Whalley and Azizi, 1994).

Profiles from the middle lobe exhibit reflectors that are slightly more discontinuous and steeply dipping. Similar to the lower lobe, reflectors in the middle lobe are more inclined near the margins of the rock glacier. The structures identified in the middle lobe
(characterized by reflectors that dip towards one another and overlain by an area lacking reflectors) are interpreted to represent thaw pits within the active layer of the rock glacier. The area overlying the dipping reflectors is thought to represent voids containing water, which would account for the lack of reflectors. These structures appear to occur randomly through the profile.

Reflectors in the profile collected from the upper lobe of the rock glacier exhibit sharp, relatively discontinuous and chaotic reflectors, in comparison to the reflectors observed in the middle and lower lobes. Previous workers have identified angular reflectors in rock glaciers and permafrost that are similar to those identified in this study (Hinkel et al., 2001 and Degenhardt and Giardino, 2003). Degenhardt and Giardino (2003) describe subsurface angular reflectors that correspond to furrows and ridges on the surface of the Yankee Boy rock glacier in Colorado. Although the resolution of the profiles in this study is less sensitive to thick layers, we suggest that the angular reflectors identified in the upper lobe of the Lehman Cirque rock glacier are also subsurface manifestations of the surface topography. The subsurface reflectors identified in the middle and lower lobes are less angular and exhibit more lateral continuity. This observation is consistent with the geomorphic expression of the surface topography between each lobe; the upper lobe exhibits the best-preserved furrows and ridges, while the middle and lower lobes exhibit moderate to weakly preserved furrows and ridges.

We interpret the presence of chaotic reflectors in the active layer of the upper lobe as evidence that it still contains lenses of segregated ice in the active layer (Moorman et al., 1994 and Hinkel et al., 2001). Alternatively, it has been suggested that irregular layers may indicate the development of less dense permafrost resulting in the deformation of
the ice and sediment (Berthlin et al., 2000). This may be the case with the Lehman Cirque rock glacier because it is located in a hot arid climate flanked to the east and west by desiccated pluvial lakes that could provide sufficient salt to increase the unfrozen water content (Berthling et al., 2000). The lack of common midpoint and semblance analysis make it impossible to evaluate whether the chaotic reflectors identified in the profiles are solely a function of segregated ice or a warm environment with high salt input, or a combination of both processes.

The layered structures identified from profiles collected perpendicular to the flow of the upper lobe of the rock glacier were also identified on the profiles collected parallel to the flow of the upper lobe. This suggests that they are spatially consistent in a three-dimensional reference frame. The resolution of our data, and lack of geospatial reference points, limits our ability to interpret the significance of the layering observed in the Lehman Cirque rock glacier. However it is likely that they are similar to features described by Isaksen et al. (2000), Berthling et al. (2001), and Degenhardt and Giardino (2003). These studies attributed the presence of layered reflectors in rock glaciers to the development of interstitial ice in response to alternating periods of snow accumulation and sediment influx followed by rapid burial. The Lehman Cirque rock glacier is located at the head of a steep, narrow, north-facing cirque, and it is composed of well-jointed quartzite that is subject to active frost action and frequent rockfalls and dry grain flows. During spring melt, sediment is transported into the cirque via steep gullies and bedrock chutes (Fig. 3.16). The morphology of the cirque and the abundance of steep, narrow chutes helps funnel debris and sediment onto the modern rooting zone of the rock glacier. A similar scenario is envisioned during periods when the Lehman Glacier was more extensive because the steep cirque walls were (and are) laterally continuous past the
terminus of the lower lobe (Fig. 3.16). As previously mentioned, the limiting factor for the development of the Lehman Cirque rock glacier was almost certainly the development and preservation of snow/ice owing to the high rates of talus input into the cirque. This process can be observed in the field during the summer months when quartzite debris topples onto the upper slope of the Lehman glacier. Therefore, the Lehman Cirque rock glacier is located in an ideal setting to develop alternating layers of ice/snow and quartzite debris. During times of more extreme diurnal temperature flux, we envision talus input is more frequent, which would contribute to the development of layered structures in the rock glacier. Thus, we interpret the lateral continuity of the layering, the consistency of the layered structures between the three lobes, and the similarity to features described in the literature, as evidence that layers of ice and sediment/debris developed through processes described by Berthling et al. (2000), Isaksen et al. (2001) and Degenhardt and Giardino (2003).

Conclusions

Geomorphic and GPR data suggest that the Lehman Cirque rock glacier is composed of at least three lobes, the development of which reflect the episodic nature of the Lehman Glacier during the Neoglacial. However, these lobes are not superimposed on one another in the traditional ‘layer cake’ stratigraphy predicted by proponents of a glacial genesis for rock glaciers. We propose a recessional genesis model for the Lehman Cirque rock glacier, in which each lobe developed at the terminus of the Lehman Glacier as it progressively decayed to its modern position. This model is based on (1) geomorphic evidence that suggests each lobe has a distinct and unique preferred orientation, and (2) the presence of deep arcuate pits at the terminus of the upper and
Figure 3.16: USGS digital orthophoto quadrangle (DOQ) draped over a triangular integrated network (TIN) that illustrates the spatial relationships between the Lehman rock glacier and the surrounding topography.
middle lobes. The presence of these pits suggests that the upper lobe of the Lehman Cirque rock glacier has not overridden the middle lobe, and that the middle lobe has not overridden the lower lobe. We also suggest the lower segment of the Lehman Cirque rock glacier is older than 1200 years BP and may represent the initiation of a 4000-year BP advance documented by Davis (1998). However, without absolute age constraints this correlation is tentative.

We suggest that the chaotic reflectors and layered structure observed in the radar profiles indicate that the upper lobe of the Lehman rock glacier retains interstitial ice and supports a periglacial genesis. The middle lobe may also contain interstitial ice; however the reflectors and layering are not as well preserved or developed. Both the middle and upper lobe contain structures thought to represent thaw pits. The lack of such features in the lower lobe suggests that either they never formed or, because the lower lobe is much older, any thaw pits that developed have since filled in.

Furthermore, we suggest that GPR may be a useful tool for calibrating relative age relationships between individual lobes on rock glaciers, based on consistency between the angularity of subsurface reflectors and the degree of surface topography.

Future Work

A GPR survey using the Pulse Ekko system with lower frequency antennas will provide more detail in the deeper section of the Lehman Cirque rock glacier. The Pulse Ekko system is suggested rather than the Noggin Plus system because 25–100 MHz are currently available for that system, and because CMP analysis can be readily obtained. Laser surveying should be conducted to obtain precise topographic constraints for topographic correction. Although the Lehman Cirque rock glacier is located in Great
Basin National Park, a convincing case should be made to park officials to permit the installation of bolts on prominent talus blocks for a long-term velocity measurement study. The upper lobe is assumed to be inactive, however velocity measurements over a 5–10 year period would test this assumption. In the context of global warming, it is imperative that we begin to monitor the activity of landforms, such as rock glaciers, that are sensitive to climate change. Finally, a first order attempt should be made to quantify the volume of the debris in each lobe and the annual debris input into the cirque during the winter and summer months to evaluate the significance of the Lehman Cirque rock glacier as a talus transport system throughout the Neoglacial.

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Abstract

We utilize ArcView GIS, a Snow and Ice Model (SIM), modern climate station data, and digital elevation models (DEMs), to predict the modern and late Pleistocene distribution of perennial snow and ice in the Snake Range, Nevada. The Snake Range exhibits evidence of extensive glaciation during the last glacial maximum. By calculating the distribution of perennial snow and ice under a variety of temperature and precipitation boundary conditions and comparing model results to observed glacial landforms, we attempted to predict likely Late Quaternary temperature and precipitation conditions for the Snake Range. The results suggest that the combination of sparse climate data and limitations of the SIM underpredict the expected distribution of both perennial snow and ice when compared with previous estimates of perennial snow distribution using ELAps and known Full Glacial limits that were delineated using moraines.

The results from the SIM also suggest that the limiting factor on the initiation and genesis of glaciers in the Snake Range is temperature and not precipitation; this finding agrees with previous paleoclimate studies in the Great Basin. However, we suggest that the presence of pluvial Lakes Spring and Maxey to the west of the Snake Range may have provided a significant source of winter precipitation that is not reflected in the floral and faunal record.

Introduction

The application of computer modeling to evaluate the late Quaternary paleoclimate of the Great Basin is limited (Orndorff, 1994; Jones and Orndorff, 1999; Plummer et al., 2000; Van Hoesen and Orndorff, 2001; Orndorff and Van Hoesen, 2001; and Negrini,
However, significant research has been conducted in areas surrounding the Great Basin (Singer, 1985; Barry et al., 1990; Bartlein, 1998; and Plummer and Phillips, 2003). This study evaluates the applicability of a Snow and Ice Model (SIM) developed by Orndorff (1994) to 25 m raster data from the Snake Range, Nevada. We compare output grids from SIM with previously calculated output grids from ELAps (Orndorff and Jones, 1999 and Orndorff and Van Hoesen, 2001) and the expected distribution of perennial ice based on known glacial limits delineated using lateral and terminal moraines (Plate I, in pocket; Van Hoesen, 2003a).

Previous work suggests that lake-atmosphere feedback systems existed for Lakes Lahonton and Bonneville during the late Quaternary (Hostetler et al., 1994). A number of pluvial lakes existed to the west of the Snake Range, and a small arm of Lake Bonneville developed to the east of the range. Through a first order, simplistic approach we estimate the possible influence of pluvial lakes on the development of glaciers and the implied correlation with global climate change during the Late Quaternary.

Study Area

The Snake Range is a north-south-trending range in east central Nevada, composed of a northern and southern section and rising to 3981 meters above sea level (masl) at its crest (Fig. 4.1). Glaciation was more extensive in the southern portion of the range during the last Glacial Maximum. Glaciated valleys in the Snake Range are composed of Paleozoic metasedimentary units, primarily Prospect Mountain Quartzite, Pioche Shale, and Pole Canyon Limestone, that were intruded by the Jurassic Snake Creek-Williams Canyon pluton (Drewes, 1958; Whitebread, 1969; Lee et al., 1970; Lee and Van Loenen, 1971; Lee et al., 1981; Lee and Christiansen, 1983 a, b; Lee et al., 1986;
Figure 4.1: Line diagram illustrating the geographic extent of the Great Basin Physiographic Province in the Western United States and the location of the southern Snake Range.

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Miller, 1995 and Miller, 1999). A discussion on the glacial history of the Snake Range is provided by Piegat (1980), Osborn (1990), Osborn and Bevis (2001), and Van Hoesen and Orndorff (2003). The Lehman Cave weather station, located at 2073 masl, records a modern MAAT of 9.59° C. The mean coldest month is January (-2.69 °C), and the warmest month is July (22.75 °C). Mean annual precipitation is approximately 33.22 cm; the greatest amount of precipitation falls in May (3.23 cm), and the least falls in July (1.63 cm). Average mean annual temperatures for the entire range varies from -8°C to 30°C depending on elevation, and average precipitation from the valleys to the high peaks region can vary from less than 12 cm to greater than 50 cm (Houghton et al., 1969).

During the Late Pleistocene the Snake Range experienced at least three separate episodes of alpine glaciation. The terminology used to describe glacial advances in the ranges of the Great Basin was developed in the Ruby Range; the younger is termed the Angel Lake advance and the oldest advance is the Lamoille (Blackwelder, 1934). These advances tentatively correlate to the Tahoe and Tioga glacial advances of the Sierra Nevada (Wayne, 1984; Osborn and Bevis, 2001). If these correlations are correct, then an age of 19,000 to 20,000 years BP can be assigned to the Lamoille advance, and an age of 14,000 to 12,000 years BP can be assigned to the Angel Lake advance.

The youngest glacial advance is thought to represent glacial activity during the Neoglacial that tentatively correlates to the Matthes Stade of the Sierra Nevada and the Gannett Peak Stade of the Rocky Mountains (Birman, 1964; Benedict, 1968; Piegat, 1980; and Osborn and Bevis, 2001). An approximate age on the Neoglacial advance is 5,000 to 8,000 years B.P. (Denton and Karlen, 1973; Porter and Denton, 1967; and Madsen and Currey, 1979). The most extensive glaciation occurred in the central part of range, primarily in the Wheeler Peak Quadrangle. However, scattered glaciers
developed in the northern and southern sections of the range (Piega, 1980; Osborn and Bevis, 2001; and Van Hoesen, 2003a).

During the Late Pleistocene, pluvial Lakes Spring and Maxey developed in Spring Valley to the west of the Snake Range, and one of the arms of pluvial Lake Bonneville formed in Snake Valley to the east of the Snake Range (Snyder and Langbein, 1962; Snyder et al., 1964; Mifflin and Wheat, 1979; Currey, 1982; and Reheis, 1999; and Oviatt et al., 1992). Figure 4.2 shows the spatial distribution of pluvial lakes in relation to ranges that experienced alpine glaciation during the Last Glacial Maximum (LGM). Significant changes in the paleoecology occurred in the Snake Range, and throughout the Great Basin, in response to the expansion of glacial and pluvial systems (Mead et al., 1982; Thompson and Mead, 1982; Mead, 1984; Mead and Mead, 1985; and Turnmire, 1985).

Methodology

We use the Snow and Ice Model (SIM) and ELÁpse 3.0 developed by Orndorff (1994) and Jones and Orndorff (1999) in combination with 25 m USGS digital elevation models (DEMs) to predict the spatial distribution of perennial snow and ice under a Full Glacial climate. DEMs are also used to calculate slope, aspect, solar insolation, and land surface curvature. Our methodology incorporates shading effects and seasonal solar insolation variability based on the sun's position at the summer and winter solstices. In addition, we use slope and curvature surfaces to estimate stability and delineate areas most suitable for snow and ice accumulation. Using the SIM, we estimate the spatial distribution of perennial snow and ice for specified climate boundary conditions.

Modern climate data were obtained from the Western Regional Climate Center to
CLIMATE STATIONS
1. Ely Yelland Field
2. Ruth
3. McGill
4. Great Basin National Park
5. Currant Highway Station
6. Lund
7. Geyser

PLUVIAL LAKES
A. Lake Meinzer (1768 m)  I. Lake Diamond (1829 m)
B. Lake Surprise (1567 m)  J. Lake Newark (1847 m)
C. Lake Lahonton (1332 m)  K. Lake Hubbs (1920 m)
D. Lake Tahoe (1926 m)  L. Lake Franklin (1850 m)
E. Lake Dixie (1097 m)  M. Lake Clover (1730 m)
F. Lake Desatoya (1889 m)  N. Lake Waring (1761 m)
G. Lake Toiyabe (1702 m)  O. Lake Spring (1759 m)
H. Lake Railroad (1402 m)  P. Lake Maxey (1762 m)

Figure 4.2: Spatial distribution of pluvial lakes relative to mountain ranges that incurred glacial activity during the Last Glacial Maximum. Climate stations used to develop climate coefficients are indicated with black circles.
calculate modern monthly temperature maximum (Tmax), monthly temperature minimum (Tmin), and monthly precipitation (Precip; Table 4.1). These data were used to create climate coefficients using a multivariate regression between elevation and latitude of each station (Table 4.2). Climate data in the Great Basin are sparse, and climate coefficients were calculated using the seven climate stations listed in Table 4.3. Tmax and Precip grids were created using a C+ program written by Orndorff (1994). The flowchart in Figure 4.3 summarizes the steps used to create these climate grids. The climate coefficients were entered into ELApse and Tmax and Precip grids were entered into the SIM so that modern temperature and precipitation can be perturbed to reflect estimated Full Glacial climate.

<table>
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<th>Longitude</th>
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Table 4.3: Climate stations used to calculate modern climate coefficients and Tmax and Precip grids used with ELApse and the SIM.

The SIM calculates the spatial extent of perennial snow and ice within the domain of the DEM and creates a raster grid that depicts the extent of predicted perennial snow and ice. It has previously been shown that a similar model, ELApse, overestimates the extent of snow and ice because it fails to incorporate the effects of topography and solar
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<th>May</th>
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Table 4.1: Summary of mean monthly Tmax, Tmin, and Precip from climate station used to calculate climate coefficients and climate grids. (EYF = Ely Yelland Field, R = Ruth, M = McGill, GBNP = Great Basin National Park, CHS = Cunant Highway Station, L = Lund, and G = Geyser).
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<th>May</th>
<th>June</th>
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Table 4.2: Monthly climate coefficients produced using a multivariate regression between altitude and elevation at each climate station.
Figure 4.3: Flowchart illustrating the steps required to create Tmax and Precip grids for the SIM.
insolation (Jones and Orndorff, 1999; Orndorff and Van Hoesen, 2001; and Van Hoesen and Orndorff, 2001). Therefore, raster overlays of slope, curvature, and shading created in ArcView to simulate alpine microclimates are used to refine the grids produced by the SIM to provide a more realistic depiction of the spatial extent of perennial snow and ice.

Results

We used the SIM to estimate perennial snow and ice for various parts of the Snake Range under modern conditions, and then we perturbed the modern climate by $-2^\circ$, $-3^\circ$, $-4^\circ$, $-5^\circ$, $-5.5^\circ$, $-6^\circ$ and $-6.5^\circ$C. Each perturbation produces an output grid depicting the distribution of perennial snow, perennial glaciers, a summary of results, and the volume of snow and ice over time. Tables 4.4 and 4.5 provide a summary of the results for each area of interest of the Snake Range (Fig. 4.4). Each portion of the Snake Range investigated in this study, with the exception of Lincoln Peak, exhibits a similar trend with increased equilibration time occurring at the boundary when perennial snow transitions to perennial ice. A similar transition occurs when perennial snow first develops on Bald Mountain, Lincoln Peak, and Williams Canyon where SIM doesn’t predict modern perennial snow. However, the change in equilibration time between bare ground and the first appearance of perennial snow is less drastic than the change observed during the transition from perennial snow to ice.

We also used the SIM to estimate perennial snow and ice for Baker Valley by perturbing the temperature by $-5^\circ$C and increasing precipitation by 1, 2, 3 and 4 cm to evaluate the effects of changing temperature versus increased precipitation. We compared the extent of perennial snow and ice produced with just a temperature change...
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<th>ΔT (°C)</th>
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<th>Perennial Snow (#cells)</th>
<th>Glaciers (#cells)</th>
<th>Snow Area (km²)</th>
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Table 4.4: Perennial snow and ice output from the SIM under modern and perturbed climate for Baker and Lehman Valley, Bald Mt and Lincoln Peak.
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Table 4.5: Perennial snow and ice output from the SIM under modern and perturbed climate in Snake Valley, Williams Canyon, and on Mt Moriah.
Figure 4.4: Location of each area of interest with respect to the Snake Range crest.
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<th>Grid Extent (#cells)</th>
<th>Perennial Snow (#cells)</th>
<th>Glaciers (#cells)</th>
<th>Snow Area (km²)</th>
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<td>5.13</td>
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</tbody>
</table>

Table 4.6: Summary of snow and ice predicted in Baker Valley with a -5°C temperature change and 1 cm incremental changes in precipitation.
of -6.5°C with the extent of perennial snow and ice created by changing temperature by -5°C and +1, +2 and +3 cm of precipitation. The results are summarized in Table 4.6.

The SIM predicts perennial snow under modern conditions for a number of regions in the Snake Range, however once the topographic controls (slope, curvature, and shading) were applied, perennial snow is only predicted with a temperature change of -5°C, and the spatial distribution of snow dramatically decreases. Perennial ice throughout the Snake Range is only predicted on high alpine ridges and summits, isolated to an elevation range of ~3400-3900 masl. The perennial ice that was predicted occurred only on steep slopes, convex ridges, or west-facing slopes. These are regions that are not characterized as suitable for the development of small alpine glaciers. Once the grids representing the optimum topographic conditions were applied, perennial ice was not present in any of the investigated areas.

**Baker Valley**

The SIM predicts the occurrence of perennial snow at the head of Baker Valley under modern conditions. Perennial snow continues to develop without the co-occurrence of ice until a temperature change of -4°C. The area of perennial snow reaches a maximum cover of 16.69 km² at -6.5°C where perennial ice reach a maximum cover of 2.6 km². There is an abrupt boundary in the equilibration run time, which drastically increases from 117 years at -3°C to 23,076 years at -4°C (Fig. 4.5).

Although perennial snow is predicted under modern conditions in Baker Valley, once the grids representing the optimum topographic controls are applied, perennial snow isn’t predicted until a temperature change of -2°C (Fig. 4.5). However, a significant area of perennial snow isn’t produced until a temperature change of -4°C.
Figure 4.5: (A) Illustration of the considerable jump in equilibration time once perennial ice begins to develop in Baker Valley, (B) predicted evolution of perennial snow in Baker Valley with temperature perturbations of -2, -3, -4, -5, -6 and -6.5°C using the SIM, and (C) predicted zones of actual accumulation derived using topographic controls.
Bald Mountain

The SIM doesn’t predict perennial snow on Bald Mountain until a temperature change of -4°C with a total area of 0.29 km². Perennial snow develops without the co-occurrence of ice until -6.5°C when the area of perennial snow reaches a maximum of 5.78 km² and perennial ice reaches a maximum cover of 0.04 km². There is an abrupt boundary in the equilibration run time, which drastically increases from 124 years at -6°C to 40,429 years at -6.5°C (Fig. 4.6).

Perennial snow isn’t predicted until a temperature change of -4°C, however once the grids representing the optimum topographic controls are applied, perennial snow isn’t predicted until a temperature change of -5°C (Fig. 4.6).

Lehman Valley

The SIM predicts the occurrence of perennial snow at the head of Lehman Valley under modern conditions. Perennial snow continues to develop without the co-occurrence of ice until a temperature change of -4°C. The area of perennial snow reaches a maximum cover of 16.49 km² and perennial ice reach a maximum cover of 2.91 km² at -6.5°C. There is an abrupt boundary in the equilibration run time, which increases from 3,026 years at -4°C to 11,394 years at -5°C (Fig. 4.7). Equilibration run time then decreases to 7,365 years at -6°C and then increases to 8,616 at -6.5°C.

Although perennial snow is predicted under modern conditions in Lehman Valley, once the grids representing the optimum topographic controls are applied, perennial snow isn’t predicted until a temperature change of -2°C (Fig. 4.7). However, a significant area of perennial snow isn’t produced until a temperature change of -3°C.

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Figure 4.6: (A) Illustration of the considerable jump in equilibration time once perennial ice begins to develop on Bald Mountain, (B) predicted evolution of perennial snow on Bald Mountain with temperature perturbations of -2, -3, -4, -5, -6 and -6.5°C using the SIM, and (C) predicted zones of actual accumulation derived using topographic controls.
Figure 4.7: (A) Illustration of the considerable jump in equillibration time once perennial ice begins to develop in Lehman Valley, (B) predicted evolution of perennial snow in Lehman Valley with temperature perturbations of -2, -3, -4, -5, -6 and -6.5°C using the SIM, and (C) predicted zones of actual accumulation derived using topographic controls.
Lincoln Peak

The SIM doesn't predict perennial snow on Lincoln Peak until a temperature change of -3°C with a total area of 0.33 km². Perennial snow develops without the co-occurrence through a temperature change of -6.5°C when the area of perennial snow reaches a maximum of 12.7 km². No perennial ice is predicted for Lincoln Peak under any climate regime using SIM. No abrupt boundaries occur in the equilibration run times, which range from 2 years to 155 years (Fig. 4.8).

Perennial snow isn't predicted until a temperature change of -3°C, however once the grids representing the optimum topographic controls are applied, perennial snow isn't predicted until a temperature change of -4°C (Fig. 4.8). However, a significant area of snow isn't predicted until a temperature change of -5°C.

Mt Moriah

The SIM predicts the occurrence of perennial snow on Mt Moriah under modern conditions. Perennial snow continues to develop without the co-occurrence of ice until a temperature change of -6°C. The area of perennial snow reaches a maximum cover of 18.57 km² and perennial ice reaches a maximum cover of 1.21 km² at -6.5°C. There is an abrupt boundary in the equilibration run time, which drastically increases from 101 years at -5°C to 5,297 years at -6°C and another from 5,297 years at -6.5°C to 397,732 years at -6.5°C (Fig. 4.9).

Although perennial snow is predicted under modern conditions on Mt Moriah, once the grids representing the optimum topographic controls are applied, perennial snow isn't predicted until a temperature change of -4°C (Fig. 4.9). However, a significant area of perennial isn't produced until a temperature change of -5°C.
Figure 4.8: (A) Illustration of the considerable jump in equilibration time once perennial ice begins to develop on Lincoln Peak, (B) predicted evolution of perennial snow on Lincoln Peak with temperature perturbations of -2, -3, -4, -5, -6 and -6.5°C using the SIM, and (C) predicted zones of actual accumulation derived using topographic controls.
Figure 4.9: (A) Illustration of the considerable jump in equilibration time once perennial ice begins to develop on Mt Moriah, (B) predicted evolution of perennial snow on Mt Moriah with temperature perturbations of -2, -3, -4, -5, -6 and -6.5°C using the SIM, and (C) predicted zones of actual accumulation derived using topographic controls.
Snake Valley

The SIM predicts the occurrence of perennial snow in Snake Valley under modern conditions. Perennial snow continues to develop without the co-occurrence of ice until a temperature change of -6.5°C. The area of perennial snow reaches a maximum cover of 6.41 km² and perennial ice reaches a maximum cover of 0.04 km² at -6.5°C. There is an abrupt boundary in the equilibration run time, which drastically increases from 148 years at -6°C to 17,756 years at -6.5°C (Fig. 4.10).

Although perennial snow is predicted under modern conditions in Lehman Valley, once the grids representing the optimum topographic controls are applied, perennial snow isn’t predicted until a temperature change of -4°C (Fig. 4.10). However, a significant area of perennial isn’t produced until a temperature change of -5°C.

Williams Canyon

The SIM doesn’t predict perennial snow in Williams Canyon until a temperature change of -3°C with a total area of 0.09 km². Perennial snow develops without the co-occurrence of ice until -6.5°C when the area of perennial snow reaches a maximum of 5.34 km² and perennial ice reaches a maximum cover of 0.01 km². There is an abrupt boundary in the equilibration run time, which drastically increases from 147 years at -6°C to 3,655 years at -6.5°C (Fig. 4.11).

Perennial snow isn’t predicted until a temperature change of -3°C, however once the grids representing the optimum topographic controls are applied, perennial snow isn’t predicted until a temperature change of -4°C (Fig. 4.11). However, a significant area of snow isn’t predicted until a temperature change of -5°C.
Figure 4.10: (A) Illustration of the considerable jump in equillibration time once perennial ice begins to develop in Snake Valley, (B) predicted evolution of perennial snow in Snake Valley with temperature perturbations of -2, -3, -4, -5, -6 and -6.5°C using the SIM, and (C) predicted zones of actual accumulation derived using topographic controls.
Figure 4.11: (A) Illustration of the considerable jump in equilibration time once perennial ice begins to develop in Williams Canyon, (B) predicted evolution of perennial snow in Williams Canyon with temperature perturbations of -2, -3, -4, -5, -6 and -6.5°C using the SIM, and (C) predicted zones of actual accumulation derived using topographic controls.
Evaporation Estimates

The area of pluvial lakes Maxey and Spring were calculated using a 25 m DEM based on the spatial extent of Late Quaternary pluvial lakes in the Great Basin (Reheis, 1999). There is little improvement over the methodology employed by Reheis (1999) who calculated area based on contour lines extracted from a 250 m DEM. A summary of the characteristics of Lakes Maxey and Spring is provided in Table 4.7.

<table>
<thead>
<tr>
<th>Lake</th>
<th>Area_R (m²)</th>
<th>Area_VH (m²)</th>
<th>Perimeter (m)</th>
<th>Max Elev (m)</th>
<th>Min Elev (m)</th>
<th>Mean Elev (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Spring</td>
<td>617849191.16</td>
<td>617837120.00</td>
<td>220693.49</td>
<td>1946.00</td>
<td>1641.00</td>
<td>1718.00</td>
</tr>
<tr>
<td>Maxey</td>
<td>210872297.15</td>
<td>210866464.00</td>
<td>71775.06</td>
<td>17590.00</td>
<td>1752.00</td>
<td>1759.00</td>
</tr>
</tbody>
</table>

Table 4.7: Summary of spatial characteristics of Lakes Spring and Maxey, located to the west of the Snake Range. (Area_R = area calculated by Reheis, 1999 and Area_VH = area calculated in this study. Max Elev = maximum elevation of lake, Min Elev = minimum elevation of lake, and Mean Elev = mean elevation of each lake).

Evaporation from each lake was estimated using Equation 1.0 from Mifflin and Wheat, (1979).

\[
\text{Evap} = 248 - 1.46 \left( \frac{\text{Elev}}{10} \right) + 0.0036 \left( \frac{\text{Elev}}{10} \right)^2
\]  

(1.0)

Where Evap = evaporation (cm/yr) and Elev = elevation (m). The vector files produced by Reheis, (1999) depicting the spatial extent of Late Pleistocene pluvial lakes were used to extract elevation values for Lakes Maxey and Spring. Equation 1 was applied to the estimated highest elevation of Lakes Maxey and Spring, producing the evaporation estimates and comparisons shown in Table 4.8.
<table>
<thead>
<tr>
<th>Technique</th>
<th>Evaporation (cm/year)</th>
</tr>
</thead>
<tbody>
<tr>
<td>EV&lt;sub&gt;sl&lt;/sub&gt;</td>
<td>Lake Spring 116.8</td>
</tr>
<tr>
<td>EV&lt;sub&gt;mw&lt;/sub&gt;</td>
<td>Lake Spring 102.57</td>
</tr>
<tr>
<td></td>
<td>Lake Maxey 102.52</td>
</tr>
</tbody>
</table>

Table 4.8: Summary and comparison of evaporation estimates for Lakes Spring and Maxey based on previously published data and techniques described by Synder and Langbein, 1962 (EV<sub>sl</sub>) and Mifflin and Wheat, 1979 (EV<sub>mw</sub>).

Discussion and Conclusions

Geomorphic evidence indicates that late Quaternary glaciers were far more extensive and developed at lower elevations than glaciers predicted using the SIM (Van Hoesen, 2003a; b). Thus, it appears the SIM underestimates the perennial distribution of perennial ice and snow in the Snake Range. In addition, the spatial distribution of perennial snow predicted by the SIM is significantly less than that predicted using ELAps (Jones and Orndorff, 1999 and Orndorff and Van Hoesen, 2001). For example, ELAps predicts perennial snow for every grid cell within the high peaks region of the southern Snake Range with a temperature change of -6°C. Once topographic controls are applied, this overestimate of perennial snow provides a more realistic representation. The distribution of perennial snow predicted by ELAps is more consistent with the expected accumulation zones of known glaciers in the investigated area (Fig. 4.12; Jones and Orndorff, 1999 and Orndorff and Van Hoesen, 2001).

The discrepancies between the expected spatial distribution of perennial snow and ice and the distribution predicted using the SIM are likely introduced by the paucity of climate data and the methodology used to generate minimum temperature values (Tmin;
Figure 4.12: Raster grids depicting the spatial distribution of perennial snow in the high peaks region of the southern Snake Range, NV. (A) distribution of snow under modern conditions, (B) distribution of snow with a -2°C temperature change, (C) distribution of snow with a -4°C temperature change, and (D) distribution of snow with a -6°C temperature change.

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There is evidence suggesting that alpine glaciers in the Great Basin reached their maximum extent at approximately ~15-18 ka and started retreating by ~13 ka (Elliott-Fisk, 1987 and Dorn et al., 1990). Lake Bonneville in particular dropped 100 m at ~14.5 ka in response to downcutting of the Zenda threshold (Oviatt et al., 1992). However, many Great Basin pluvial lakes didn’t reach their maximum extent until approximately 14 ka. This suggests that glaciers started retreating prior to the maximum extent of pluvial lakes in the Great Basin (Antevs, 1925; Lajoie, 1968; Scott et al., 1993; Wayne, 1984; and Bevis, 1995). This relationship is consistent with the hypothesis that following the northward transition of the jet stream and storm tracks, in response to reduced continental ice sheets, winter months became warmer, cloudier, and experienced less precipitation (Antevs, 1948; Kutzbach and Guetter, 1986; Benson and Thompson, 1987; COHMAP Member, 1988; Allen and Anderson, 1993 and Kutzbach et al., 1993). Decreased precipitation, decreased evaporation and increased temperature would facilitate the decay of glaciers. This decay would be accompanied by increased fluvial output and the subsequent growth of pluvial lakes to the east of the Schell Creek and Snake Ranges in Spring and Snake Valley respectively.

However, it has been suggested that evaporation from pluvial lakes could provide an ancillary source of moisture for the development of alpine glaciers in the Great Basin (Hostetler et al., 1994; Bevis, 1995; Oviatt, 1996; 1997; Hostetler et al., 2000). Estimated evaporation rates for Lakes Maxey and Spring are likely overestimates since rates of evaporation rates are depressed in proportion to salinity, water color and depth, and wind velocities (Harbeck, 1955; Langbein, 1961; Turk, 1970; and Smith and Perrott, 1983). Evaporation is also strongly controlled by fluctuating lake levels, humidity, air and surface water temperature and variations in cloud cover (Benson, 1981; 1986).
These parameters were not evaluated in this study so evaporation rates represent rough estimates based solely on a regression between evaporation, surface area and elevation.

The results of climate perturbations in the Snake Range suggest the limiting factor on the development of perennial snow and ice is temperature and not small changes in precipitation. This inference is consistent with previous paleoclimate studies in the Great Basin (Sears and Roosma, 1961; Wells, 1983; Thompson, 1992; Bevis, 1995; and Hostetler and Clark, 1997). However, the presence of Lakes Spring and Maxey directly west of the Snake Range could have provided a significant source of moisture (~120 cm/yr) facilitating the development of alpine glaciers. Much of this moisture would be lost during evaporation, changes in wind direction, and interception by vegetation. If the Snake Range intercepted just 5-10% of this moisture during the winter months, this could potentially represent a 6-12 cm increase in precipitation. This rough estimate doesn’t take into account the increase in precipitation of 20 cm and a reduction in evaporation by 21.5 cm in order to develop and sustain Lake Spring suggested by Snyder and Langbein (1962). It also doesn’t account for the possible input of moisture from a number of pluvial lakes located to the west of the range (Fig. 4.2; Reheis, 1999). Although, it is likely both Lake Spring and Maxey were partially covered by ice similar to Lake Lahonton and Bonneville (Benson, 1981 and Hostetler et al., 1994). The discrepancy between suggesting an increase in precipitation in the Snake Range derived from pluvial lakes and previous paleoclimate and paleoecology studies that suggest the region only experienced a small increase in precipitation, may result from the seasonality of the precipitation. Increased precipitation may not be reflected in the flora or faunal distribution in the Snake Range if it falls during the winter months and not during the
early spring or growing season. During warmer months, much of the moisture derived from pluvial lakes would be lost to the atmosphere rather than falling as precipitation.

Oviatt (1996, 1997), who suggests that changes in pluvial Lake Bonneville were synchronous with Heinrich and Bond events, further enhanced the interrelationship between pluvial lakes, the atmosphere, and alpine glaciers identified by Hostetler et al. (1994). If the hypothesis that increased winter moisture derived from pluvial lake systems facilitated the growth and preservation of alpine glaciers in the Great Basin is valid, then it is plausible the decay of glaciers prior to the onset of lake degradation created a lag in the response time of pluvial lakes to the northward migration of the jetstream during Heinrich and Bond events. Benson (1981) and Smith (1979) suggested the meltwater derived from glacial ice in the Sierra Nevada were not sufficient to sustain pluvial Lakes Lahonton and Searles. However, due to the smaller area of Lakes Spring and Maxey, increased runoff and fluvial output could delay the collapse of pluvial lake systems. Thus, in some cases the synchronicity of falling lake levels with Heinrich Events may be reflecting a synchronicity with the final decay of alpine glaciers and the loss of fluvial input.

Future Work

Numerous mountain ranges throughout the Great Basin experienced extensive glaciation during the Last Glacial Maximum (LGM). Almost every glaciated range was associated with a pluvial lake or lakes. The estimated potential winter evaporation and precipitation relationship needs to be evaluated for each of these ranges to test for consistency throughout the Great Basin. It is possible that glaciation in many of these ranges developed under increased winter precipitation drawn from local pluvial lake
systems. However, a higher resolution estimate of lake evaporation based on mean
monthly air temperature, monthly means of air temperature, air humidity and solar
radiation should be calculated. This can be accomplished by using the Complementary
Relationship Lake Evaporation (CRLE) model that is widely used by electricity
companies in Brazil (Hostetler et al., 1990; dos Reis and Dias, 1998; and Leydecker and
Melack, 2000). CRLE provides estimates of monthly evaporation and the change of
stored enthalpy in lake waters and equilibrium temperature, a proxy for the actual water
surface temperature.

This study suggests the SIM drastically underestimates the spatial distribution of
perennial snow and ice, likely caused by the technique used to derive Tmin. However,
ELApsce appears to be a more accurate tool for calculating perennial snow distribution.
Therefore, the next step would involve incorporating the plastic C+ subroutine from the
SIM (i.e. – the technique used to initiate the flow of ice from one grid cell to another)
into the ELApsce code to better evaluate this technique on higher resolution digital
elevation models. Our goal is to create an extension for the program ArcGIS so users
can more readily access the integrated features of ELApsce and the SIM.

To truly evaluate the interrelationship between falling lake levels and deglaciation in
the Snake Range, and throughout the Great Basin, a higher resolution absolute age
chronology must be established. Van Hoesen and Orndorff (2003) suggest age
constraints in the Snake Range will be difficult to obtain because of the lack of material
suitable for radiocarbon dating, cosmogenic surface exposure dating and luminescence
dating. However, the Schell Creek Range contains a similar glacial record as the Snake
Range and may provide opportunities to further constrain the glacial chronology in both
ranges. Similarly, the shorelines of Lakes Spring and Maxey have not been dated or
investigated for paleoclimatic indicators. Investigating the relationship between the Schell Creek Range and Lake Spring and Maxey may provide insight into the paleoclimate of the Snake Range and possibly illuminate the relationship between pluvial lake systems and alpine glaciers.

Acknowledgements

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References


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Van Hoesen, J.G., 2003, Surficial geology of glaciated canyons in Windy Peak an Wheeler Peak Quadrangles, Southern Snake Range, Nevada. This Volume: Plate I.


CHAPTER 5

A COMPARATIVE SEM STUDY ON THE MICROMORPHOLOGY OF GLACIAL AND NON-GLACIAL CLASTS WITH VARYING AGE AND LITHOLOGY: EXAMPLES FROM THE ARCTIC AND SOUTHWESTERN UNITED STATES.

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Abstract

Preliminary research on striated clasts from a variety of depositional environments suggests that scanning electron microscopy (SEM) of striated clasts of varying lithology in diamictons may prove useful in distinguishing a glacial origin. The evaluation of SEM analysis of clasts from diamictons is an applicable technique to define a glacial origin requires a better understanding of the micro-features on glacial and non-glacial clasts. We describe the micromorphology of surface textures and characteristics for samples of quartzite, granite, limestone, basalt, chert, pillow basalt, and quartz pebbles collected from a variety of depositional environments. This study suggests that certain micro-features are useful for differentiating glacial and non-glacial deposits based on the micromorphology of entrained clasts.

Introduction

Limited data have been published on the analysis of micro-features and characteristics of striated clasts thought to have a glacial origin (Wentworth, 1926; Judson and Barks, 1961; Bjørlykke, 1974; Hicock and Dreimanis, 1989). Most research concerning the micromorphology of tills has focused on individual quartz grains (Brown, 1973; Krinsley and Doornkamp, 1973; Folk, 1975; Mahaney et al., 1988; Mazzulo and Ritter, 1991; Mahaney et al., 1991; Campbell and Thompson, 1991; Mahaney and Kalm, 1995; Mahaney, 1995; Mahaney et al., 1996; Mahaney and Kalm, 2000). Surface textures of quartz grains have been used to determine local ice transport vectors, ice thickness, and depositional environment. SEM results from these studies established that sediment affected by glacial transport and deposition exhibits distinct microtextural characteristics (Table 5.1). However, these characteristics are not isolated to
<table>
<thead>
<tr>
<th>Feature</th>
<th>References</th>
</tr>
</thead>
<tbody>
<tr>
<td>Abrasion features</td>
<td>Mazzullo and Ritter, 1991</td>
</tr>
<tr>
<td>Arc shaped steps</td>
<td>Campbell and Thompson, 1991; Mahaney et al., 1996, 2001; Mahaney and Kalm, 2000; Mahaney, 1995</td>
</tr>
<tr>
<td>Breakage blocks</td>
<td>Campbell and Thompson, 1991</td>
</tr>
<tr>
<td>Chattermarks</td>
<td>Campbell and Thompson, 1991; Folk, 1975</td>
</tr>
<tr>
<td>Curved grooves</td>
<td>Campbell and Thompson, 1991; Mahaney et al., 1988, 1996, 2001; Mahaney, 1995; Mahaney, 1990b</td>
</tr>
<tr>
<td>Crescentic gouges</td>
<td>Campbell and Thompson, 1991; Mahaney et al., 1988, 1996, 1995; Watts, 1985</td>
</tr>
<tr>
<td>Crushing features</td>
<td>Mahaney, 1990a; Mahaney, 1990b; Mahaney et al., 1988</td>
</tr>
<tr>
<td>Deep troughs</td>
<td>Mahaney et al., 1996, 2001; Mahaney and Kalm, 2000</td>
</tr>
<tr>
<td>Edge rounding</td>
<td>Campbell and Thompson, 1991; Mahaney et al., 1996, 2001; Mahaney and Kalm, 2000; Mahaney, 1995; Mahaney et al., 1991; Mahaney et al., 1988</td>
</tr>
<tr>
<td>Fracture faces</td>
<td>Mahaney et al., 2001; Mahaney and Kalm, 2000; Mahaney et al., 1996</td>
</tr>
<tr>
<td>Lattice shattering</td>
<td>Mahaney et al., 1996, 2001; Mahaney and Kalm, 2000; Mahaney, 1995</td>
</tr>
<tr>
<td>Linear steps</td>
<td>Mahaney et al., 1996, 2001; Mahaney and Kalm, 2000</td>
</tr>
<tr>
<td>Star-cracking</td>
<td>Campbell and Thompson, 1991</td>
</tr>
<tr>
<td>Subparallel crushing planes</td>
<td>Mahaney, 1995; Mahaney et al., 1988</td>
</tr>
</tbody>
</table>

Table 5.1: Summary of previously described surface textures and features identified on quartz grains and bedrock using a scanning electron microscope.
glacially derived quartz grains. While previous research has established the merits of SEM analysis of quartz grains in glacial research, this study suggests that SEM analysis of larger clasts in diamictons may also prove useful in defining a glacial origin. We hypothesize that, clasts larger and commonly softer than quartz grains, exhibit some of the same microtextural characteristics if the deposit from which they were collected has a glacial origin.

The goals of this study are to (1) document the various micro-scale textures and surface features exhibited by clasts of differing lithologies from glacial environments, (2) compare and contrast the observed textures and features with features identified on clasts from non-glacial environments, and (3) establish whether SEM analysis of striated clasts in diamictons is an applicable tool to infer a glacial origin.

Methods

Striated clasts ranging in size from 5.0 cm to 0.75 meters were collected from known glacial, eolian, fluvial, and lacustrine environments throughout the southwestern United States and Canada during 2000 to 2002. Till samples from Allan Hills, Antarctica were also provided by Phil Holme. This study focused on the southwest and high latitude regions because we reason that preservation of surface features will increase with decreased physical and chemical weathering typical of these regions. Two sample sets from glacial and non-glacial deposits were chosen for analysis: those that displayed visible surface striations or markings and those that lacked visible surface features. More samples were collected in glacially affected environments because one of the goals of this study is to characterize the microfeatures related to glaciation. Samples collected from
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Figure 5.1: Location of study sites located in the Yukon, Antarctica and the southwestern United States.
## Table 5.2: Summary of sample sites, sample ID designation, geographic coordinates, lithology, relative age (EP = Early Pleistocene, MP = Middle Pleistocene, LP = Late Pleistocene, T = Tertiary, H = Holocene), and depositional environment.

<table>
<thead>
<tr>
<th>Name</th>
<th>ID</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Lithology</th>
<th>Relative Age</th>
<th>Environment</th>
</tr>
</thead>
<tbody>
<tr>
<td>Glacial</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Yosemite NP, CA</td>
<td>DP</td>
<td>37° 50' 39.7&quot;</td>
<td>119° 27' 31.3&quot;</td>
<td>Granite</td>
<td>MP</td>
<td>Bedrock</td>
</tr>
<tr>
<td>Devils Postpile NM, CA</td>
<td>YR</td>
<td>64° 01' 37.7&quot;</td>
<td>139° 21' 45.2&quot;</td>
<td>Chert</td>
<td>~1.5MA</td>
<td>Till</td>
</tr>
<tr>
<td>Fort Selkirk, YT</td>
<td>SM</td>
<td>36° 16' 13.6&quot;</td>
<td>115° 40' 25.6&quot;</td>
<td>Limestone</td>
<td>EP</td>
<td>Till</td>
</tr>
<tr>
<td>Spring Mountain's, NV</td>
<td>SC</td>
<td>38° 15' 53.8&quot;</td>
<td>114° 15' 51.8&quot;</td>
<td>Quartzite</td>
<td>EP</td>
<td>Moraine crest</td>
</tr>
<tr>
<td>Snake Creek, NV</td>
<td>AH-1</td>
<td>62° 48&quot;</td>
<td>137° 20&quot;</td>
<td>Quartz</td>
<td></td>
<td>Till</td>
</tr>
<tr>
<td>Allan Hills, Antarctica</td>
<td>AH-2</td>
<td>62° 48&quot;</td>
<td>137° 20&quot;</td>
<td>Pillow Basalt</td>
<td>Oligocene?</td>
<td>Till</td>
</tr>
<tr>
<td>Fish Lake, UT</td>
<td>FLO</td>
<td>38° 35' 27.4&quot;</td>
<td>111° 40' 33.9&quot;</td>
<td>Trachyte</td>
<td>EP</td>
<td>Moraine crest</td>
</tr>
<tr>
<td></td>
<td>FL</td>
<td>38° 34' 15.5&quot;</td>
<td>111° 41' 55.5&quot;</td>
<td>Trachyte</td>
<td>LP</td>
<td>Moraine crest</td>
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<td>La Sal Mountains, UT</td>
<td>Tuku</td>
<td>38° 26' 55.1&quot;</td>
<td>109° 15' 26.9&quot;</td>
<td>Trachyte</td>
<td>LP</td>
<td>Rock glacier lobe</td>
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<td>Bonneville Shoreline, UT</td>
<td>LB</td>
<td>41° 36'34&quot;</td>
<td>113° 37'02&quot;</td>
<td>Quartzite</td>
<td>MP</td>
<td>Wave cut shoreline</td>
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<tr>
<td>Klondike Gravel, YT</td>
<td>KG</td>
<td>64° 01' 37.7&quot;</td>
<td>139° 21' 45.2&quot;</td>
<td>Chert</td>
<td>LP</td>
<td>Glacial outwash</td>
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<tr>
<td>North Fork Baker Creek, NV</td>
<td>NFB</td>
<td>38° 43'05&quot;</td>
<td>114° 23'31&quot;</td>
<td>Quartzite</td>
<td>H</td>
<td>Fluvial</td>
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<tr>
<td>Spring Mountain's, NV</td>
<td>SMFC</td>
<td>36° 27'20&quot;</td>
<td>115° 67'04&quot;</td>
<td>Limestone</td>
<td>H</td>
<td>Fluvial</td>
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Mono Basin age deposits in the Sierra Nevada based on clast preservation and post-depositional modification of the landform (Russell, 1889; Sharpe and Birman, 1963; and Yount and LaPointe, 1997). However, the lack of suitable organic material for radiocarbon analysis precludes accurate age determination. Approximately 30 limestone clasts were collected from the till and four limestone clasts were collected from fluvial deposits upvalley from the deposit during the summer of 2000.

**Southern Snake Range**

The Snake Range is located in east central Nevada within the boundaries of the Basin and Range physiographic province. It has long been recognized that the Snake Range was glaciated during the Late Quaternary (Gilbert, 1875; Russell, 1884; Weldon, 1956; Kramer, 1962; Whitebread, 1969; Osborn, 1990; Piegat, 1980; Osborn and Bevis, 2001). Two discrete intervals of glaciation in the Snake Range are correlative with Lamoille and Angel Lake advances (Sharpe, 1938; Wayne, 1983; Osborn and Bevis, 2001). The Angel Lake advance is approximately correlative with the Tioga stage, and the Lamoille advance with the Tahoe stage as described in the Sierra Nevada. The glaciated portions of the Snake Range are primarily early Paleozoic Prospect Mountain Quartzite, Pioche Shale, and Pole Canyon Limestone intruded by the Jurassic age Snake Creek – Williams Canyon Pluton (Drewes, 1958; Whitebread, 1969; Lee et al., 1970; Lee and Van Loenen, 1971; Lee et al., 1981; Lee and Christiansen, 1983 a, b; Lee et al., 1986; Miller, 1995 and Miller, 1999). Two quartzite clasts were collected from the crest of a dissected Lamoille moraine (Fig. 5.2b) in Snake Creek at approximately 2587 masl and two quartzite clasts were collected from Holocene fluvial deposits in North Fork Baker Creek at approximately 3038 masl during the summer of 2001.
Figure 5.2: Field photos illustrating typical environment of selected samples. (a) Big Falls Till in the Spring Mountains, NV, (b) Lamoille (Tahoe) age moraine in Southern Snake Range, (c) view of rock glacier on Mt Tukuhnikivatz from a lateral moraine, (d) polished surface of granite in Yosemite NP, (e) Upper surface of polished basalt columns in Devils Postpile NM, (f) moderately consolidated diamicton along Yukon River, (g) upper surface of Klondike Gravel SW of Dawson City, and (h) wave polished quartzite shoreline in NW Utah.
La Sal Mountains, Utah

The La Sal Mountains are located in east–central Utah approximately 35 km east of Moab, Utah. Evidence for Pleistocene glaciation includes characteristic U-shaped valleys, moraines, and striated and polished bedrock (Richmond, 1962; Nicholas, 1994). There is also evidence of Holocene glaciation in at least one locality in the range (Nicholas, 1991; Nicholas and Butler, 1996). The range is composed of Late Cretaceous to early Tertiary hypabyssal intrusions of trachyte and rhyolite porphyries overlain by a thick Quaternary mantle (Hunt, 1958; Richmond, 1962; Ross, 1998). Two samples were collected from a moraine dissected by Brumley Creek, a tributary to Pack Creek, on the southwest flank of Mount Tukuhnikivatz (3805 masl; Fig. 5.2c).

Fish Lake Plateau, Utah

The Fish Lake Plateau is located southeast of Salt Lake City, Utah. The plateau is composed of Tertiary volcanics, primarily red to purple trachyte and light gray sanidine trachyte (Hardy and Muessig, 1952). Tertiary sedimentary rocks underlie the volcanic successions and these in turn are thought to overlie the Green River formation (Hardy and Muessig, 1952). Two episodes of glaciation are described as Wisconsin I and Wisconsin II that more than likely represent the Tioga/Tahoe and/or Angel Lake/Lamoille glacial stages. Samples were collected from a younger Wisconsin II terminal moraine at the mouth of Pelican Canyon and an older Wisconsin I terminal moraine north of Pelican Bay (also referred to as Widgeon Bay).

Yosemite National Park

Yosemite National Park is located in the west–central portion of the Sierra Nevada batholith. The eastern section of Yosemite Valley is composed of granites intruded by the Late Cretaceous Tuolumne Intrusive Suite (Bateman, 1992; Ratajeski et al., 2001).
Matthes (1930) recognized the development and preservation of two ages of polished surfaces indicative of older and younger glacial advances in Yosemite Valley. Four samples of younger polished surfaces of Late Cretaceous granite were collected near Tenaya Lake during summer 2001 (Fig. 5.2d).

**Devils Postpile National Monument**

Devils Postpile National Monument (DPNM) is located in southern California east of Yosemite National Park and southwest of Mammoth Lakes. This flow is composed of Pliocene columnar trachybasalt – trachyandesite whose upper surface was scoured and polished during the last glaciation (Fig. 5.2e; Bailey, 1987 and Bailey and Hill, 1987). Four samples of basalt were collected from the upper surface of the polished basalt flow during the summer of 2001.

**Yukon Territory**

A chert pebble was collected from a 1.5Ma year old (Ma) diamicton along the Yukon River downstream from Fort Selkirk during summer 2001 (Fig. 5.2f; Jackson et al., 2001). During the summer of 2002, six chert pebbles were collected from the upper unit of the Klondike Gravel/Conglomerate southwest of Dawson City (Fig. 5.2g) and two chert pebbles were collected from Pleistocene age glaciofluval gravel at the head of Dominion Creek (Froese, et al., 2000 and Froese, 1997).

**Allan Hills, Antarctica**

In the Allan Hills region, extensive peperites, hyaloclastites and irregular intrusive masses are intermixed with tuffs and tuff-breccias of the Mawson Formation (Ballance et al., 1971). These rocks, and correlatives 1500 – km along strike in the Transantarctic Mountains, form regional pyroclastic deposits transitional between Beacon Supergroup terrestrial sedimentation and Ferrar Supergroup flood basalt volcanism (Ballance et al.,
1971; Denton and Hughes, 2002; and Atkins et al., 2002). Five quartz pebbles and two pebbles of pillow basalt were extracted from a sample of the Sirius Group till, provided by Phil Holme. The Sirius Group most likely developed during the Oligocene to Miocene, although the age of this deposit has been controversial (Bruno et al., 1999; Stroeven and Klehman, 1999; Holme, 2001; and Denton and Hughes, 2002).

**Lake Bonneville Shoreline, Utah**

Lake Bonneville occupied closed depressions in the eastern Great Basin and covered approximately 20,000 square miles at its greatest extent. Lake Bonneville created three major shorelines, the Provo, Bonneville, and Stansbury shorelines that can be identified as terraces on many mountains throughout western Utah, eastern Nevada, and southern Idaho (Oviatt et al., 1992). The Provo shoreline has an approximate age of 16,800 – 16,200 years BP, the Bonneville has an approximate age of 18,000 – 16,800 years BP, and the Stansbury shoreline has an approximate age of 24,400 – 23,200 years BP (Oviatt et al., 1992). 4 samples of Quartzite of Clarks Basin were collected from a Bonneville - age terrace in northeastern Utah at approximately 1553 masl (Fig. 5.2h).

**Results**

Previous studies have focused on scanning electron microscopy of quartz grains from known glacial deposits or active glacial environments. These studies documented and described distinctive micro-scale features that are characteristic of glacial environments. The samples in this study exhibit many of these glacially derived features depending on lithology, age, and source of the clast. Of particular interest is the presence of arc-shaped and linear steps, and conchoidal and sub-parallel crushing features assumed to develop through compression and grinding during glacial transport. Specific SEM-
observed micro–features for each lithology and depositional environment are summarized in Figure 5.3.

**Spring Mountains**

Limestone clasts from the Big Falls till display both macro-scale features and micro-scale features characteristic of a glacial origin. The surfaces of most clasts are strongly polished and contain striations (1.0 – 4.0 mm wide), chatter marks (3.0 – 5.0 mm), crescentic gouges (1.0 – 3.0 cm), and rare small–scale flute topography (1.0 – 3.0 cm amplitudes) that are all visible to the unaided eye. Chatter marks occur individually on the surface of the clasts, within the striations, and they commonly occur in small lenses of micrite in the limestone. Scanning electron microscopy of fourteen striated limestone clasts (with c–axes that varied from about 10 cm to 30 cm in length) reveals distinctive micro-scale features attributed to the effects of glaciation including the following: straight grooves (Fig. 5.4a), crescentic gouges, arc-shaped steps, curved grooves, conchoidal and subparallel crushing features, and linear steps. Straight grooves range in width from 2.0 – 10 µm, crescentic gouges range in size from 75 – 100 µm wide with variable length, arc-shaped steps are 20 – 100 µm wide, and crushing features are 5.0 – 20 µm wide. None of the samples exhibit scouring features or v-shaped percussion features, and very few contain features that appear to have been rounded since formation. Many show evidence of dissolution etching, and some have weathered surfaces. In contrast, limestone clasts sampled from fluvial deposits exhibit rounding, visible scouring features and lack linear striations. SEM analyses indicate the presence of well–developed v-shaped percussion features that are ~100 – 175 µm wide at their apex, prominent scouring features, and edge rounding (Figs. 5.7a – b).
## Micro Features Observed On Striated Samples

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<tr>
<th>Sample Location</th>
<th>AH-1</th>
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<th>FLO</th>
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Figure 5.3: Summary of observed microfeatures observed on striated samples from various depositional environments and lithologies. Sample identification summary: AH = Allan Hills, DP = Devils Postpile, FL = Fish Lake, KG = Klondike Gravel, LB = Lake Bonneville, NFB = North Fork Baker, SC = Snake Creek, SM = Spring Mountains, SMF = Spring Mountains Fluvial, TUKU = La Sal Mountains, YR = Yukon River, and YS = Yosemite National Park.
Devils Postpile

The upper surface of the Devils Postpile basalt flow is strongly polished and striated (Figs. 5.4g – h). Striations ranging in width from ~1 – 3 mm and crescentic gouges ~1 – 1.5 cm wide are visible with the unaided eye. Under SEM magnification, these samples exhibit sub-parallel linear fractures, straight grooves that are 1 – 2 µm wide, linear steps that are 25 – 75 µm wide, deep troughs that are 5 – 15 µm wide, conchoidal crushing features that are 10 – 90 µm wide, and arc-shaped steps ~50 – 100 µm wide. The glacial polish that is so obvious in the field is easily identified in the SEM as relict polished, although this surface does exhibit evidence of dissolution etching.

Fish Lake

Two ages of sanidine trachyte were collected from the Fish Lake region representing the older Early Pleistocene and younger Late Pleistocene glacial advances. The surface of the older sample is more weathered and lacks visible striations, while the younger sample is well weathered but retains visible striations ranging in width from 1 – 4 mm (Fig. 5.5a – b). Viewed with the SEM, the older sample exhibits arcuate grooves that range in width from 75 – 125 µm, arc-shaped steps ranging in width from 50 – 80 µm, and crushing features. Relict polish occurs as pre-weathered surfaces that are weathered and eroded. Arcuate grooves and arc-shaped steps are weakly preserved in some instances but difficult to recognize. The younger sample exhibits arcuate grooves that range in size from 50 – 100 µm wide, arc-shaped steps ranging from 30 – 70 µm wide, sub-parallel crushing planes that are 10 – 20 µm wide, fracture faces, relict polish, and crushing features (Figs. 5.5c – d).
Figure 5.4: SEM photomicrographs of representative glacial features from Spring Mountain and Devil's Postpile samples. (a) Polished limestone surface showing numerous deep troughs (arrows) and arcuate grooves/striations (ag), (b) deep arcuate groove surrounded by surface with some evidence of dissolution etching, (c) lunate fracture containing conchoidal crushing features (ccf), (d) close-up of crushing features, (e) subparallel crushing features (crf) and breakage blocks (bb), (g) relict polished surface on basalt, (h) close-up of polished basalt showing numerous arcuate grooves.
Snake Range

Two quartzite samples collected from the crest of a Lamoille age moraine exhibit well-preserved striations ~ 2 – 4 mm wide, visible with the unaided eye. Under magnification, these samples contain numerous well-preserved features, including sub-parallel crushing features that are 10 – 20 μm wide, conchoidal fractures 2 – 5 μm wide, sub-parallel linear fractures, straight grooves that are 5 – 10 μm wide, linear steps that are 5 – 30 μm wide, fracture faces, curved grooves that are 20 – 30 μm wide, arc-shaped steps that are 10 – 20 wide, chattermarks, and star cracking (Figs. 5.5e – h).

In contrast, two quartzite samples collected from a fluvial deposit are sub-rounded to rounded and are heavily scoured. Viewed under the SEM, they exhibit well-developed, v-shaped percussion features that are ~150 – 180 μm wide at their apex, moderately developed scouring features, and minor edge rounding.

La Sal Mountains

Trachyte clasts collected from the surface of a Late Pleistocene moraine exhibit well-developed striations. These samples display crushing features 15 – 30 μm wide, curved grooves that are 10 – 20 μm wide, and arc-shaped steps that are 15 – 40 μm wide (Figs. 5.6 a – b). They also show evidence of mechanically upturned plates, dissolution etching, and relict polish preserved as pre-weathered surfaces.

Allan Hills, Antarctica

Two lithologies were identified in the Allan Hills till sample. Five quartz pebbles ranging in size from 1 – 3 cm and two pebbles of pillow basalt ranging in size from 0.5 to 1.0 cm were analyzed. The quartz pebbles were well rounded and finely polished, and the pillow basalt was sub-rounded and heavily weathered. The quartz pebbles exhibit a combination of glacial and non-glacial characteristics including v-shaped percussion and...
Figure 5.5: SEM photomicrographs illustrating typical features and textures on Fish Lake (old and young) and Snake Creek samples. (a) relic polished surface (pws) with deep arcuate grooves, (b) close-up of deep arcuate groove containing crushing features, (c) quartz crystal showing fracture faces (ff) and subparallel crushing features (spcf), and (d) another quartz crystal exhibiting fracture faces, subparallel crushing features and conchoidal crushing features (ccf), (e) sub-parallel conchoidal fractures, conchoidal fractures and crushing features, (f) close-up of conchoidal fractures and crushing features, (g) conchoidal fractures and subparallel crushing features, and (h) fracture faces with sub-parallel conchoidal fractures and conchoidal crushing features.

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scouring features and arc-shaped steps. V-shaped percussion and scouring features are the most common, but well-developed arc-shaped steps and rare lunate fractures are also present (Figs. 5.6c – d). The v-shaped percussion features range in size from 10 to 100 \( \mu m \), arc-shaped steps range in size from 50 to 100 \( \mu m \), and lunate fractures are 80 – 125 \( \mu m \) wide. The pillow basalt pebbles lack diagnostic micro-features that could be used to identify an erosional origin. They are strongly weathered and exhibit high relief and abundant abrasion features.

**Yosemite National Park**

Samples were collected from strongly polished surfaces exhibiting small (~0.5 – 1.0 mm) striations visible with the unaided eye. Samples are moderately weathered and have high mineral relief. SEM analysis reveals the presence of 2 – 5 \( \mu m \) wide, sub-parallel linear crushing features, straight grooves that are 2 – 8 \( \mu m \) wide, curved grooves that are 50 – 150 \( \mu m \) wide, arc-shaped steps that are 30 – 60 \( \mu m \) wide, deep troughs that are 25 – 75 \( \mu m \) wide, breakage blocks and fracture faces (Figs. 5.6e – f). The polished surfaces visible in the field are well preserved at the micron scale as pre-weathered relict surfaces with little evidence of dissolution etching. These samples are characterized by high relief and sharp angular features with no evidence of v-shaped percussion cracks or scouring features.

**Yukon Territory**

A chert sample collected from a ~1.5 Ma diamicton along the Yukon River is well polished and well rounded and exhibits small (0.5 – 1 mm wide) visible striations and chatter marks. SEM analysis of the clast shows the presence of straight grooves that are 5 – 15 \( \mu m \) wide, deep troughs that are 10 – 15 \( \mu m \) wide, arc-shaped steps that are 10 – 20
Figure 5.6: SEM photomicrographs of representative glacial features on clasts from the La Sal Mountains, Antarctica, Yosemite NP, and Yukon River. (a) Surface of polished quartz pebble showing numerous v-shaped percussion features (arrows) and rare arcuate gouges (ag), (b) close-up of v-shaped percussion feature (vpf), (c) close-up of lunate fracture (lf) surrounded by smaller v-shaped percussion features (arrows), (d) surface of Alan Hills pillow basalt showing no evidence of preserving micro-features, (e) quartz crystal exhibiting deep arcuate groove on a relict polished surface (pws), (f) close-up of arcuate groove and relict polish, (g) sub-parallel straight grooves, (h) sub-parallel linear grooves and arcuate lunate fracture or gouge.
μm wide, rare v-shaped percussion features, evidence of dissolution etching, and some scouring features (Figs. 5.6g – h).

Chert samples from the Klondike Gravel are well polished and round to sub-rounded. Small striations and abrasion features are visible in the field using a hand lens. SEM analyses indicate the presence of well-developed, v-shaped percussion features that are ~25 – 100 μm wide at their widest point, prominent scouring features, and edge rounding (Figs. 5.7c – d).

Lake Bonneville Shoreline

Samples from a Bonneville-age, wave-cut terrace are well polished and show visible signs of scouring and abrasion. Striations, chattermarks, and crescentic gouges are not visible in the field with a hand lens. These samples exhibit v-shaped percussion features that are ~25 – 75 μm wide, arcuate grooves that are ~ 5 – 20 μm wide, scouring features, and relict polish preserved as pre-weathered surfaces (Figs. 5.7e – f).

Discussion

Mahaney and Andres (1996) identified conchoidal and linear fractures on quartz grains from fluvial and eolian environments. They noted the depth and preferred orientation of fractures and striations on grains affected by glacial processes. The consistency of the preferred orientation, and the well-developed morphology of such features suggest they are diagnostic of a glacial origin and are not derived from eolian or fluvial processes. The development of some microfeatures has also been attributed to dissolution etching in weak hydrochloric acid, and the effects of river ice (Wentworth, 1928 and Dunn, 1933). The morphology of these striations and polish, and abundance of clasts from the same environment exhibiting similar characteristics suggest they are not
Figure 5.7: SEM photomicrographs of representative non-glacial features on fluvial and lacustrine samples from the Spring Mountains, Klondike Gravel, and Lake Bonneville shoreline. (a) Surface of limestone showing numerous v-shaped percussion features, (b) close-up of scouring features and relict surface, (c) numerous v-shaped percussion features and rare arcuate gouges, (d) close-up of v-shaped percussion feature outlined in 7c, (e) surface of shoreline exhibiting scouring features, arcuate gouges but predominantly v-shaped percussion features, and (f) shoreline exhibiting numerous v-shaped percussion features.
dissolution features. It is also unlikely they reflect an environment that could sustain river ice since almost all of these samples are collected from arid regions or alpine localities. However, in this study we differentiate between three groups of samples, (1) those that were only affected by glacial activity, (2) those only affected by non-glacial processes, and (3) samples that were transported or modified by both glacial and non-glacial processes. Each group of samples exhibits distinct microfeatures indicative of the specific environment from which they were derived (Fig. 5.8).

Microfeatures are most readily preserved in samples with a homogeneous mineralogy. Samples of polymineralic granite, trachyte, and pillow basalt display the most evidence for post-depositional weathering and weakly preserved microfeatures compared to more homogeneous samples of chert, quartzite and limestone. Samples from the Snake Range and Spring Mountains (SC and SM) exhibit the best-preserved features but are thought to be older than many of the other samples. It is likely the Snake Creek samples retain their features because quartzite is so resistant to erosion and weathering, although the Spring Mountain samples are anomalously pristine given their presumed age. Limestone is highly susceptible to dissolution etching and weathering, so the excellent preservation of microfeatures on samples from the Big Falls till is thought to reflect their burial through mass wasting events and subsequent isolation from most physical and chemical weathering processes (Orndorff and Van Hoesen, in press). However, the older and younger Fish Lake samples illustrate the expected effects of weathering on the preservation of microfeatures through time. Features observed on the older Fish Lake (FLO) samples are not only more subdued and less well preserved than the younger Fish Lake (FLY) samples, but many of the characteristic features attributed to glacial action.
Figure 5.8: Frequency of observed microfeatures differentiated by glacial, non-glacial or mixed formational environments.
are lacking. The preservation of these features is directly affected by the amount of time spent on the surface of a landform or contained in a glacial/non-glacial deposit. Samples from Devils Postpile, Yosemite Valley, Snake Creek, Fish Lake, and La Sal Mountains were collected at sites with direct exposure to the elements. Increased surface area and exposure to physical and chemical weathering processes most certainly causes a higher rate of microfeature degradation than typically observed on quartz grains. The likelihood of preservation for microfeatures as a function of relative age and lithology is illustrated in Fig. 5.9.

V-shaped percussion features were identified only on clasts affected by glaciofluvial activity (Krinsley and Doornkamp, 1973) with the exception of AH – 1 and WR. Previous studies have documented the prevalence of v-shaped percussion features on quartz grains from tills deposited by alpine glaciers compared to continental glaciers (Mahaney et al., 1988 and Mahaney, 1995). These samples (AH – 1 and WR) were collected from tills deposited by continental glaciers and were likely influenced by either fluvial or glaciofluvial processes prior to incorporation into glacial ice. The lack of v-shaped percussion features on samples collected from bedrock surfaces is consistent with our expectations; however the absence of such features on samples from alpine moraines and rock glaciers suggests a short transport distance. Field relationships suggest the source for these samples is primarily frost-riven talus from local cirques (e.g. samples from a lateral moraine crest, terminus of a rock glacier breaching a cirque bedrock sill). We infer that talus is transported into the cirque basin through mass wasting and rockfall events and that transport downvalley is primarily achieved through glacial advance rather than fluvial or glaciofluvial processes, so the lack of v-shaped percussion features is not surprising.
Many of the microfeatures (arc-shaped steps, crescentic gouges, deep grooves, etc.) described in this study are thought to develop during transport by glaciers greater than 500 m thick in response to glacial abrasion and high basal shear stress (Mahaney et al., 1988; 1991; and Mahaney, 1990b). This relationship is not reflected in this study because many of the samples are less competent than quartz grains and therefore should require less basal shear stress to produce similar features. They may also be produced in alpine environments that exhibit steep gradients and coarser material entrained in glacial ice that can affect the average shear stress at the base of the glacier (Benn and Evans, 1998). We suggest the most important factor is the competence and homogeneity of bedrock and the material being transported by the glacier. The homogeneity of samples is not thought to affect the frequency of occurrence of microfeatures as much as the preservation of these features.

Conclusions

This study indicates that distinct microfeatures are present on striated clasts with a glacial origin and that they can be used to differentiate between diamictons of a glacial versus non-glacial origin (Fig. 5.8). Arc-shaped steps, chattermarks, curved grooves, deep troughs, fracture faces, linear steps, relict polish, subparallel linear fractures, and subparallel crushing planes dominate samples from the seven glacial environments. These features were primarily observed on samples with a known glacial origin. The characteristic microfeatures for the samples with a non-glacial origin are breakage blocks, scouring features, and percussion features. The samples that were affected by both glacial and non-glacial processes exhibit a combination of these microfeatures including
Figure 5.9: Schematic diagram illustrating the most likely relationship between the preservation of microfeatures in relation to lithology and age of the samples. With increasing age and decreasing lithologic competence microfeatures should be less common, however, the samples in this study with the best preserved microfeatures are assumed to be older than many of the other samples.
arc-shaped steps, chattermarks, and v-shaped percussion features. As with previous studies addressing the micromorphology of quartz grains, we conclude that the presence of arc-shaped steps, chattermarks, deep grooves and arcuate and conchoidal crushing features are diagnostic indicators of a glacial origin, and v-shaped percussion marks and scouring features are characteristic of fluvial and glaciofluvial environments. Further study on additional depositional environments and lithologies is necessary to evaluate the consistency of microfeatures. Furthermore, additional work can also be done on the correlation between specific features documented on striated clasts and the estimated ice thickness.

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CHAPTER 6

SUMMARY OF MAJOR RESULTS

The papers presented in this dissertation attempt to evaluate the Late Quaternary environment(s) of the Snake Range, east central Nevada. Geomorphic evidence indicates the range supported both glacial and periglacial climate regimes; more than likely, although a tentative hypothesis, they evolved concurrently and may have supported one another through positive feedback mechanisms. Previous workers have described much of the glacial geology in the Snake Range, however an absolute age chronology is lacking. Plate 1 presents the efforts of a surficial mapping project in the Wheeler Peak and Windy Peak quadrangles and provides a clearer representation of the glacial geology with respect to Quaternary and Holocene deposits. The spatial distribution of both glacial and periglacial landforms provides paleoclimatic information derived using field evidence and computer modeling. Although the modeling predicts less snow and ice than required to form the observed glacial moraines, the evolution and distribution of snow and ice under perturbed climate is consistent with modern alpine environments. Analyses of these data suggest, similar to previous authors (Sears and Roosma, 1961; Thompson and Mead, 1982; Wells, 1983; Dohrenwend, 1984; Thompson, 1992; Bevis, 1995; and Hostetler and Clark, 1997), that temperature played a more significant role in the development and growth glaciers in the Snake Range. However,
there may be a relationship between winter moisture input from pluvial lakes and glaciation that has yet to be fully evaluated in the interior Great Basin.

Periglacial Landforms

Without a higher resolution inventory of periglacial features and a better understanding of their spatial relationships to documented glacial deposits, our understanding of Late Quaternary conditions throughout the interior Great Basin will be limited. This study presents new data documenting the distribution of relict periglacial features and provides evidence suggesting the Snake Range supported discontinuous permafrost during the LGM and/or Late Quaternary.

The presence of rock glaciers, solifluction lobes, garlands, sorted and unsorted polygons and circles and cryoplanation terraces in both the northern and southern Snake Range suggests snow cover was thin enough to allow discontinuous permafrost to develop. Barsch (1978) and Jakob (1992) state that rock glaciers define the lower limit of discontinuous permafrost. Therefore, the presence of five rock glaciers in the Snake Range suggests discontinuous permafrost was present in the Snake Range during the Late Quaternary. The distribution of permafrost in the range is consistent with this inference since patterned ground is only found at elevations higher than each rock glacier. However, without absolute age constraints, it is impossible to tell whether they developed coeval with the maximum extent of glaciation or whether they developed during deglaciation as temperatures increased and snow cover decreased. The fact that almost all patterned ground is found on flat lying surfaces, are moderately to well developed features, and that glaciers developed in cirques directly east of all four major cryoplanation terraces, suggests they could have developed during the LGM with
sufficient westerly winds blowing snow out of the saddles into adjacent cirques. This scenario could create a positive feedback in which permafrost develops as snow cover is thinned, glaciers develop as snow is blown into niches and hollows on the eastern flank of the range, and as glaciers develop they create local microclimate that help support relatively small patches of permafrost (with the exception of The Table). Therefore, it is quite likely that relict patterned ground observed in the Snake Range is considerably older than the rock glaciers.

Without digging trenches, finding exposed subsurface features like the stream cut on Mt Moriah, or using ground penetrating radar (GPR), it is difficult to address the specific evolution of most periglacial features. At best, I can infer that most features developed as ground ice accumulated in response to cooling temperatures, through freeze-thaw action, and under the influence of gravity; such as in the case of solifluction lobes and garlands. Similarly it is unclear whether stone polygons and circles and cryoplanation terraces identified in the range developed solely under the action of freeze-thaw or if solifluction, gelideflation, slope, and debris supply played an important role as suggested by Boch and Krasnov (1994), Troll (1994), and Francou et al. (2001). However, lithology and slope are certainly important factors influencing the development of certain landforms because solifluction lobes and garlands are only found on slopes steeper than ~25° and sorted polygons and circles are most strongly developed in Prospect Mountain Quartzite (PMQ). Although the platy nature of PMQ is more conducive to over steepening and downslope flow than sub-angular limestone or rounded granite, the observed lithologic control may also be related to the fact that the highest portion of the range are almost entirely composed of PMQ. Unfortunately, the geneses of many periglacial features are still contentious and poorly understood, including rock glaciers.
Proposed Genesis of Lehman Rock Glacier

The proposed genesis for the Lehman rock glacier (LRG) is not an attempt to settle the debate over the glacial versus periglacial origin of rock glaciers discussed in Chapter 2. Rather, I suggest a new genesis for tongue shaped rock glaciers through the coalescence of smaller lobate rock glaciers that developed in response to the retreating Lehman Glacier. The geomorphology and GPR findings suggest the LRG is composed of three distinct individual lobes and that it is likely the upper, and possibly the middle lobe, retain lenses of relict periglacial ice. Therefore, I suggest the LRG is an ice cemented rock glacier that developed through the action of periglacial processes. I am not suggesting that all tongue shaped rock glaciers develop through this process, but only offer an alternative periglacial model of growth. The presence of interstitial ice is significant because it suggests other rock glaciers in the range as well as throughout the Great Basin may still contain remnant ice (although not necessarily interstitial) that could be valuable sources of water in a region so desperate for water resources. However, the lack of absolute ages, surface velocity measurements and detailed descriptions of the internal structure of the rock glacier prevent a more thorough interpretation of the genesis and may possible inhibit future finite element modeling that could provide further insight into its development. Although, as I suggested in Chapter 2, an estimated age of 1200 AD to 5000 BP is a plausible assumption. The presence of the LRG and other rock glaciers is also useful for evaluating paleotemperature conditions during the Neoglacial.
Paleoclimatic Implications of Glacial and Periglacial Landforms

Osborn and Bevis (2001) provide a detailed description of glacial features throughout the Great Basin, including the Snake Range, but little detail regarding the paleoclimatic conditions in the range during the Late Quaternary is provided by the current literature. Therefore, Chapter 1 evaluates the paleoclimatic significance of glacial and periglacial deposits in the Snake Range using techniques described by Harris (1980, 1981a, b); Meierding, (1982); Greenstein (1983); Locke, (1990); Benn and Evans, (1998); Frauenfelder and Kääb (2000); Hoelzle (1996); and Frauenfelder et al. (2001).

Neoglacial temperature depression estimates calculated using modern freezing and thawing indices for relict rock glaciers, range from –0.25 °C to –1.00 °C while temperature estimates calculated using the methodology described by Frauenfelder and Kääb, (2000) and Frauenfelder et al., (2001) are 0.35 to 0.97 degrees lower than those calculated using freezing and thawing indices. The calculated mean of –1.42 °C for the higher estimates are more consistent with previously published Neoglacial temperature estimates for the Great Basin (Wayne, 1984). This suggests that rock glaciers are indeed sensitive indicators of climate change. Therefore, with the proposed global temperature increase of 1.5 to 5°C (Watson et al., 1990; Wilson and Mitchell, 1987; Schlesinger and Zhao, 1989; Knox, 1991; and Mitchell et al., 1995), active rock glaciers throughout the world should be carefully monitored for ice content and potential water loss to evaluate how quickly they respond to climate change.

By comparing modern and Full Glacial MAAT in the Snake Range, calculated using regional lapse rates, I suggest the Snake Range experienced an average MAAT depression of approximately –5.16°C to –6.61°C. Not only do these estimates of full glacial MAAT lowering compares favorably with other estimates for this region.
(Chapter 1; Table 5.9), they are also similar to the MAAT depression of −4.55°C to −5.77°C that I calculated from reconstructed Angel Lake equilibrium line altitudes (ELAs). Temperature estimates from both relict permafrost features and ELA depression are also consistent with previously published temperature gradients (Bevis, 1995). Bevis, (1995) calculated MAAT depressions of −8.0°C in the Ruby Mountains northwest of the Snake Range, −5.0°C in the Deep Creek Mountains, located to the northeast of the Snake Range, and −4.0°C in the Tushar Mountains in south-central Utah.

Possible Application of SEM for Environmental Reconstruction

One component of this study identified microfeatures on quartzite clasts from the Snake Range are similar to features observed on quartz grains that are distinctly glacial in origin (Van Hoesen and Orndorff, 2003). The results of this study suggest it may be possible to use striated clasts and surfaces to reconstruct ice thickness and debris transport mechanisms similar to Mahaney et al. (1988) and Mahaney (1990 and 1991). However, I did not attempt to evaluate these parameters because only 3 samples of quartzite were collected from the Snake Range. More collections and further study is required on the factors influencing the formation these microfeatures. Previous work has only investigated quartz grains; therefore it is unclear how microfeatures on less competent lithology will vary with ice thickness and transport histories. So although I did not use microfeatures to reconstruct Late Quaternary environmental conditions, it presents new data that suggests it may be possible with further investigation.
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DESCRIPTION OF

Holocene

**Hac** Holocene avalanche chutes: steep gullies containing frost riven debris that lies stratigraphically on Quaternary talus, glacial till and unconsolidated, poorly sorted un-differentiated alluvium. The debris is clast supported and consists of angular to subangular dark brown pebbles and boulders of Prospect Mountain Quartzite (PMQ). Surface clasts are predominantly fresh and unweathered, indicative of recent activity.

**Hrg** Holocene rock glaciers: deposits of unsorted, angular frost shattered boulders and cobbles. These deposits occur as both tongue and lobate-shaped morphologies, composed of quartzite debris that can be differentiated into an upper and lower layer. The upper layer is clast supported and consists of angular to subangular, reddish brown cobbles and boulders of PMQ. The lower layer is primarily matrix supported and consists of orange to brown, angular to subangular sand and cobbles. Rock glaciers are differentiated from protalus ramparts by the presence of furrows and ridges and the surface morphology of the terminus (Barsh, 1996).

**Hrp** Holocene protalus ramparts: deposits of unsorted, angular frost shattered boulders and cobbles of PMQ. The upper surface of these features is subdued and typically orange to dark brown.

Quaternary

**Qt2** Quaternary talus (younger): deposits of angular to sub-angular quartzite debris derived from frost riving and rapid gravity transport on steep slopes and cirque headwalls, gullies and avalanche chutes. They typically form cones or aprons that lie at or near the angle of repose along valley walls and often transition into the margins of rock glaciers. Surface clasts are light grey to brown and lie stratigraphically above older Qt1 deposits.

**Qt1** Quaternary talus (older): deposits of angular to sub-angular quartzite debris derived from frost riving and rapid gravity transport on steep slopes and cirque headwalls, gullies and avalanche chutes. They typically form cones or aprons that lie at or near the angle of repose along valley walls and often transition into the margins of rock glaciers. Surface clasts exhibit more patina, are dark brown and lie stratigraphically below younger Qt2 deposits.
DESCRIPTION OF MAP UNITS

Quaternary glacial till (Angel Lake): poorly sorted, poorly stratified deposits containing angular to subangular pebbles, cobbles and boulders of PMQ, Pioche Shale (PS), Pole Canyon Limestone (PCL), and Snake Creek-Williams Canyon pluton (SCWP). The surface morphology is hummocky and is covered with sporadic cobbles and boulders of weakly weathered PMQ and moderately weathered cobbles and boulders of SCWP. Quartzite and granite clasts are weakly faceted and exhibit rare striations.

Quaternary glacial till (Lamoille): poorly-sorted, poorly-stratified deposits containing angular to subangular pebbles, cobbles and boulders of PMQ, and Snake Creek-Williams Canyon Pluton (SCWP). The surface morphology is weakly hummocky and much more subdued than Qga deposits. A weakly developed soil mantles the surface of these deposits and surface cobbles and boulders of PMQ and SCWP are moderately to strongly weathered. This unit is correlated to Lamoille deposits described by Sharpe, 1938 rather than Pre-Angel Lake (Osborn and Bevis, 2002).

Quaternary glacial tarn or kettle: glacially-derived lakes that formed through direct glacial scouring or melting. The floor of kettle lakes are mantled with cobbles and boulders of reddish orange PMQ and SCWP.

Quaternary hummocky talus: deposits of angular to sub-angular cobbles and boulders of PMQ derived from frost riving and rapid gravity transport on steep slopes and cirque headwalls, gullies and avalanche chutes. They have a hummocky, furrowed surface morphology and were likely affected by periglacial activity (permafrost?) and occur downslope from rock glacier deposits and at the base of steep slopes. Surface clasts are light grey to brown.

Quaternary frost-riven debris: extensive deposits of frost-riven PMQ that mantles numerous slopes and tables above treeline. Quartzite debris is angular and light grey to brown with evidence for periglacial activity (cryoturbation, cryoplanation, solifluction, patterend ground and garlands).

Mesozoic/Paleozoic

Figure 1: Tongue-shaped rock glacier (Hrg) complex located below Wheeler Peak. Note Qt1 and Qt2 coalescing along the lateral margins of the rock glacier.

Figure 2: Characteristic Qt1 talus aprons. Note the lobate rock glacier in the foreground and reddish brown debris.

Figure 3: Characteristic Angel Lake till (Qga) located below Teresa Cirque. Note the angularity of the clasts, lack of soil and reddish brown coloring.

Figure 4: Characteristic Lamoille till (Qgl) in Lehman Creek. Note the more subangular soil and light grey to brown coloring.

Figure 5: Characteristic hummocky till (Qht) located in Lehman Cirque. Note the furrows and ridges of the rock glacier in the foreground.

Figure 6: Characteristic frost-riven mantle on the southern face of Bald Mountain.
stic Qt_1 talus aprons overlying Qt_2:
: glacier in the foreground exposing s.

stic Lamoille till (Qgl) located along
te the more subangular edges, a weak
brown coloring.

stic frost-riven mantle located on the
Mountain.

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Plate 2: Flowchart depicting the order of operations and the required data when running the Snow and Ice Model.