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Environmental Records in a High-Altitude Low-Latitude Glacier, Sierra Nevada, California

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**ENVIRONMENTAL RECORDS IN A HIGH-ALTITUDE
LOW-LATITUDE GLACIER,
SIERRA NEVADA, CALIFORNIA**

By

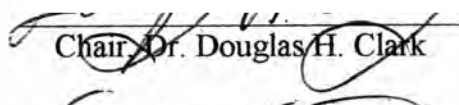
Alison J. Gillespie

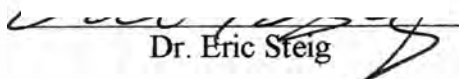
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of the Requirements for the Degree
Master of Science

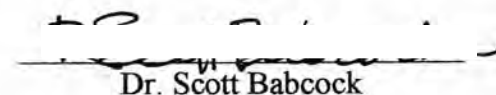
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**ENVIRONMENTAL RECORDS IN A HIGH-ALTITUDE
LOW-LATITUDE GLACIER,
SIERRA NEVADA, CALIFORNIA**

A Thesis
Presented to
The Faculty of
Western Washington University

In Partial Fulfillment
Of the Requirements for the Degree
Master of Science

by
Alison J. Gillespie
February 2006

MASTER'S THESIS

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Abstract

Glaciers are sensitive and detailed recorders of changes in local, regional, and global climate. Stable isotope variations reflect seasonal temperature changes, physical stratigraphy relates to net precipitation, and trace element concentrations mark seasonal surfaces in the ice. Previous studies have largely focused on polar ice cores because cold, dry conditions in such locations are ideal for preserving ice stratigraphy. Conversely, low-latitude alpine glaciers have been generally ignored because warmer, wetter conditions in them have been presumed to destroy or obscure the original stable isotope stratigraphy. Here, I evaluate the potential to obtain viable environmental records from an ice core of Palisade Glacier Sierra Nevada, California. A shallow, 6-m ice core drilled in the accumulation zone in late summer 2003 was analyzed at ~10-cm intervals for stable isotopes (O^{16}/O^{18} and D/H) and trace element concentrations. Five stable-isotope maxima and minima couplets occur in the ice core, varying from -16.7 to -14 ‰ ($\delta^{18}O$) and -133 to -111 ‰ (δD). The isotope values follow the global meteoric water line, indicating minimal post-depositional disruption of the ice. Maxima in six representative trace-elements (Al, Pb, Mn, V, Ba, Sr) correspond well to those in the isotope records. The covariance of stable-isotopes and trace-elements indicate that the original chemical stratigraphy is preserved in the ice core. These results show that the ice core contains at least 5 years of ice accumulation. Analyses of excess Pb and of precipitation and temperature records from a nearby weather station indicate these 5 years most likely accumulated between 1992 and 1998, and that annual layer thicknesses average about 1 m. This interpretation, combined with the substantial dust band at the surface of the ice core indicate that the five years of accumulation after 1998 (all of which were below-average snowfall) were lost to ablation. Thus, despite five years of net ice loss at the site due to surface melt, the isotopic and elemental stratigraphy is still preserved. The important implication of this finding is that ice in such temperate alpine glaciers appears to be a more robust archive of paleoclimatic information than previously recognized. Furthermore, although the ice core retrieved from the Palisade Glacier likely contains significant gaps, retrieval of a core to bedrock will likely provide the most complete ice core record of paleoclimate data available in the Sierra Nevada.

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I. Introduction

Location

The Palisade Glacier, with an area of 1.3 km², is the largest of 497 recognized glaciers in the Sierra Nevada (Trent, 1983), and the lowest-latitude (37° N), high-altitude glacier in the continental United States. The glacier extends from 3700 - 4025 m above sea level in a northeast-facing cirque along the Sierra Nevada crest in the John Muir Wilderness Area (Figure 1). Accumulation is dominated by direct snowfall, but also includes snow avalanched onto the glacier from peaks on the crest as high as 4345 m a.s.l. (Heald, 1946). The primary drainage fed by the Palisade Glacier is the North Fork Big Pine River, which empties into Owens Valley.

Palisade Glacier displays many characteristics common to alpine glaciers, including a distinct bergschrund, crevasse field, glacier tables, and moraines (Heald, 1946). The moraine at the terminus is ice-cored, with steep inner slopes and loose, lichen-free, unsorted rock debris. The most recent glacial advance in the Sierra Nevada is the Matthes (100-600 cal yr BP), which is part of the Little Ice Age (LIA) (Clark and Gillespie, 1997). Prior to the Matthes, the only glaciation recorded in the cirque moraines was the Recess Peak advance, which ended shortly before 11,190 ±70 ¹⁴C years BP (Clark and Gillespie, 1997). The presence of ice within the Palisade Glacier moraine identifies it as a relict of the LIA.

The Palisade Glacier shows clear signs of retreat. Distinct bergschrunds separate the glacier from the headwall, which contains several large couloirs and neve patches. Palisade Glacier historically connected to the northern adjoining cirque to form a more

extensive double-tongue glacier; currently the ice in the northern cirque is disconnected by a bedrock buttress and only a heavily debris-covered ice tongue remains. A melt pond at the toe of Palisade Glacier, measured at 18-22 m deep (Trent, 1983), has grown significantly since the middle of the century. The over-steepened Matthes-age ice-cored moraine below the glacier documents the greater ice extent during the Little Ice Age (LIA) (Clark and Gillespie, 1997). The modern regional snowline at 37° N latitude, as estimated by Flint (1957), is ~ 4500 m. More recently, Burbank (1991) estimated the regional snowline as ~4000 m asl near Palisade Glacier, which is above the crest. However, the firn line is clearly visible on Palisade Glacier (Figure 2) at ~3930 m a.s.l., which is ~570 m below Flint's estimate, and close to Burbank's estimate. The preservation of Palisade Glacier is probably due to several factors, including the northeast-facing aspect, shading from the steep cirque walls, and additional accumulation from avalanching snow (e.g. Clark et al, 1994).

There have been few measurements to constrain estimates of ice depth and flow on the Palisade Glacier. The first detailed topographic baseline survey was completed in 1946 (Heald, 1946). Trent (1983) measured flow velocities and distributions in 1976 by placing 15 sheet metal plates onto the glacier's surface and surveying plate positions a year later. Six plates survived, and moved at a maximum rate of 7 m/yr. The more rapidly moving center of the glacier exhibits true plastic flow (Trent, 1983), and basal slip remains undetermined. The total thickness of the glacier is also unknown, but is estimated by Trent at 100-150 m based on the melt pond depth of 18-22 m and presence of crevasses.

Weather systems in the Sierra Nevada are complex but predominantly originate from the southwest as a series of cells as opposed to a large coherent front (Krouse and Smith, 1973). Warm, moist maritime air rises orographically over the Sierra Nevada, and meets cooler, drier air from the northwest. This convergence produces the main cause of precipitation in winter months. Mean annual precipitation near the Palisade Glacier is ~100 cm water equivalent/year (we/yr), decreasing to ~25 cm we/yr at the range front to the east (Danskin, 1998). The long-term high-altitude weather station closest to Palisade Glacier is at the Central Sierra Snow Laboratory near Donner Pass (39°19'N, 120°22'W, 2100 m asl). Typical annual precipitation there is 1.3 m we (10.4 m average snowfall, with a maximum snow depth of 2.4 m we).

Ice Cores

Ice cores can be excellent sources of high-resolution paleoclimate information (Bradley, 1999). Over the past several decades, researchers have extracted and analyzed cores from around the globe, including Antarctica (Vostok, Law Dome, Byrd and Fuji Dome cores), Greenland (Camp Century, GISP2/GRIP, and Dye-3 cores), and arctic Canada (Agassiz and Devon cores). Most of these studies have focused on large ice caps and ice sheets at high latitudes that preserve long (10^6 yrs), continuous records of isotopic (oxygen and hydrogen) and other geochemical variation.

Recent coring efforts have also focused on small temperate ice masses at lower latitudes (e.g., Thompson et al., 1995). At present, six high-altitude, low-latitude ice cores have been retrieved to bedrock in Peru, Bolivia, and China (Bradley, 1999). These efforts may help fill significant geographic gaps in paleoclimate data used to test predictions of

local and synoptic climate fluctuations. The proximity of low-latitude glaciers to highly populated areas of the world make them particularly relevant in light of attempts to quantify potential human effects on the global weather cycle. The preservation and reliability of paleoclimate indicators (such as stable-isotope stratigraphy) within temperate ice, however, are not well tested. Previous studies suggest that detailed information may be obtainable from at least some of these glaciers (e.g., Naftz et al., 2002; Steig et al., 1998). Here I describe the analysis of a short ice core from the Palisade Glacier and assess the potential for obtaining climate records from longer cores from small temperate glaciers in the future.

Stable Isotopes

Stable (non-radiogenic) isotopes have been used to study moisture transport and temperatures since the late-1950's (Friedman, 1958; Epstein, 1967; Dansgaard and Tauber, 1969; Arnason, 1969; Krouse, 1970). The primary isotope ratios studied are $^{18}\text{O}/^{16}\text{O}$ and $^2\text{H}/^1\text{H}$ (deuterium), both expressed as per mil (‰) differences in a sample (spl) relative to a standard (Standard Mean Ocean Water, or SMOW: Faure, 1986, p 431).

$$\delta^{18}\text{O} = \left[\frac{(^{18}\text{O}/^{16}\text{O})_{spl} - (^{18}\text{O}/^{16}\text{O})_{SMOW}}{(^{18}\text{O}/^{16}\text{O})_{SMOW}} \right] * 10^3 \quad (1)$$

$$\delta\text{D} = \left[\frac{(D/H)_{spl} - (D/H)_{SMOW}}{(D/H)_{SMOW}} \right] * 10^3 \quad (2)$$

The notations $\delta^{18}\text{O}$ and δD thus represent the ratios of $^{18}\text{O}/^{16}\text{O}$ and $^2\text{H}/^1\text{H}$. The mass differences between the isotopes cause changes in their ratios during phase transformations; that is, fractionation patterns (degree of isotopic change during phase changes) are affected by the phase of the water and under what conditions the phase

originated (temperature, prior composition, travel path, etc.) This fractionation allows isotopic composition patterns to be used as a proxy for the climate conditions that created them (Faure, 1986, p 436).

Dansgaard (1964) documented the relationship between δD and $\delta^{18}O$ by comparing mean annual values for several sites in the Northern Hemisphere:

$$\delta D = 8.1\delta^{18}O + 11\text{‰} \quad (3)$$

This relationship has been further refined, and the equation in use today is commonly known as the global meteoric water line, expressed as,

$$\delta D = 8\delta^{18}O + 10\text{‰} \quad (4)$$

This relationship indicates that the evaporation and precipitation ratios of $\delta^{18}O$ and δD are linearly related under normal equilibrium conditions (Craig, 1961, in Faure, 1986, p 435).

Therefore, any deviation from that line may indicate disturbance of the initial isotopic ratios related to post-depositional processes such as diffusion, partial melting or sublimation, or percolation. The offset value in equation 4 (10 ‰) defines what is known as deuterium excess (d), where

$$d = \delta D - 8\delta^{18}O \quad (5)$$

Under non-equilibrium conditions, the mean value of d will deviate from 10 ‰, which provides additional information about conditions during evaporation from the moisture source. This includes mixing of surface waters by high winds, reduction of evaporation rates from high humidity, or decreases in water temperature (Bradley, 1999).

Seasonal changes in temperature are reflected in precipitation by changes in δD and $\delta^{18}O$. Due to variations in sources, atmospheric conditions, and individual storms, the relationship between stable-isotope ratios and air temperature varies with time and

location (Krouse and Smith, 1973). The relative relationship between air temperature and stable-isotope ratios reveals periods of warm and cool air temperatures even if the absolute air temperature is not directly related to a specific stable-isotope ratio. Warm precipitation has less negative $\delta^{18}\text{O}$ values, and cooler precipitation has more negative $\delta^{18}\text{O}$ values. The variation in $\delta^{18}\text{O}$ seasonal maxima and minima can be as much as 15 ‰ (Johnsen et al., 1972; Schwikowski et al., 1999), but more typical results are around 5 ‰ (Petit et al., 1997; Dansgaard et al., 1993) or even only 2 or 3 ‰ (Pohjola et al., 2005; Steig et al., 1998).

Previous work in comparing ice cores and snow-pit profiles from temperate glaciers around the world shows a definite “damping” or homogenization effect of the stable-isotope variation with depth (Blue Glacier, WA- Sharp et al., 1960; Langjökull Glacier, Iceland-Arnason, 1969; C.S.S.L.-Krouse and Smith 1973; Vernagtferner Glacier, Austria- Oerter et al., 1985; Baishui #1 Glacier, Mt. Yulong, China-He et al., 2002). This damping is thought to occur due to isotopic exchange between water vapor and firn when the firn density is $<0.55 \text{ g cm}^{-3}$, and from diffusion of water molecules in the ice when the density is greater. However, once firn becomes dense ice, homogenization of the stable-isotope stratigraphy due to molecular diffusion may take thousands of years (Bradley, 1999).

Studies during the mid-1960's of $\delta^{18}\text{O}$ values of Sierran snowpacks at the Central Sierra Snow Laboratory (C.S.S.L.) near Donner Pass emphasize the complex nature of stable-isotope interpretation in the Sierra Nevada. Meltwater is an important stratigraphic modifier, but high humidity at the surface generally prevents ^{18}O enrichment from evaporational loss. Krouse and Smith (1973) discovered that isotopes from samples

collected at C.S.S.L. did not relate directly to temperature as predicted by global trends (equations 4 and 5), nor did they retain original isotopic compositions throughout the melt season. Subsequent detailed studies at the same location show that the snowpack tends to homogenize isotopically from winter to spring (Unnikrishna, 2002) due to isotopic exchange with meltwater percolating through the snowpack. However, homogenization is not necessarily complete: for example, vertical variations of $\delta^{18}\text{O}$ in the glacier Austre Okstindbreen in Norway are not completely lost after the summer melt season (Yuanqing et al., 2002). In this instance, re-freezing hinders meltwater percolation, thus limiting exchange and preserving the record of annual winter temperatures.

The main constraint on using $\delta^{18}\text{O}$, δD and d values to correlate to specific air temperatures in the Sierra Nevada is the complexity of factors contributing to stable-isotope ratios within precipitation. It appears that the original isotopic signature of snow is spatially and temporally variable, independent of any temperature variations during storms. Snow from the edges of the small, localized cells typical of Sierran weather systems may be isotopically heavier than snow from the middle portions because of localized evaporation. This initial variability complicates and limits interpretation of the $\delta^{18}\text{O}$, δD and d values measured from snowpack profiles, independent of post-precipitation homogenization.

Trace Elements

Various workers have investigated trace-element or ionic species concentrations in glaciers and snowpacks in many locations, including Antarctica, Greenland, Asia and

the North American continent (Dick and Peel, 1985, He et al., 2002, Hinkley, 1974, Hinkley, 1993). The composition and timing of rock dust deposition helps determine seasonal surfaces, dust source areas, and air quality conditions. Trace-element concentrations include material from natural sources such as rock dust and contributions from air pollution. Studies in Antarctica suggest that elements are incorporated into snowfall in the same proportion as they are incorporated into air by winds (Dick, 1984).

The behavior of rock dust and heavy elements such as Pb within seasonal snowpacks influences interpretations based on trace-element concentrations. In south-central Alaska, trace metals in the Kahiltna and Ruth glaciers near Denali National Park were analyzed (Hinkley, 1993) and compared with results from Thompson Canyon (northernmost Yosemite National Park, west of Bridgeport CA) in the Sierra Nevada (Hinkley, 1974). Three alkali metals (K, Rb, and Cs) and three alkaline Earth metals (Ca, Sr, and Ba) were chosen to represent rock dust in each case. The Alaskan rock-dust compositions in the snowpack were interpreted to indicate annually shifting source areas. In contrast, the Sierran snowpack samples reflect a ferromagnesian composition, with ratios similar to those for a wide range of crustal rocks (Hinkley, 1993, p 20,540). The detailed variations found in Alaska were missing, and the source area for the Sierra Nevada appears to be broad and continuous.

Some common trace elements exhibit behavior different than those representing rock dust. Hinkley (1993) found that Pb levels in both the Alaska and Sierra Nevada samples are relatively high in seasonal snowpacks compared to rock sources, and are decoupled (independent) from the sources, timing, and deposition mechanisms associated with rock dust. Partial-year samples from snowpacks in Alaska and Sierra Nevada show

independently varying differences in the relative concentrations of Pb and rock dust compared with whole-year samples. For example, Pb does not experience dry deposition (i.e., direct from the atmosphere) on surfaces during summer months, unlike other components of dust, which accumulate most dramatically during dry deposition rather than during wet (snowfall) deposition. Therefore, an analysis of concentrations using trace elements representative of rock dust along with concentrations of Pb should distinguish between these modes of deposition. The “rock dust” trace elements indicate summer surfaces, whereas other trace element concentrations (such as Pb) can give information about the ice core age (J. McConnell, pers. comm., 2004).

The relationship between stable-isotope and rock-dust profiles can indicate how much post-depositional alteration has occurred. Analysis of snow pit samples collected at different times within a summer season on a temperate glacier near the Urumqi River, Tien Shan, China, indicates that the ionic species profile (representing rock dust) loses stratigraphic clarity after deposition at a faster rate than the $\delta^{18}\text{O}$ profile, creating a discrepancy between the stable-isotope and dust stratigraphy (Hou and Qin, 2002). These effects are attributed to post-depositional processes such as meltwater leaching, which preferentially uptakes ionic species over stable-isotopes.

Temperature and Precipitation

The relationship between glacier mass balance and climate is complex. Several variables affect seasonal net ablation, including but not limited to precipitation, air temperature and humidity, wind scour, avalanches, topography and aspect, crevassing, surface albedo, and cloud cover. Small changes in summer weather can have significant effects on glacier mass balances. For instance, just one or two summer snowstorms can

reduce ablation more by increasing albedo than adding mass (dry snow albedo averages 84%, while bare ice averages 30% or less). Thus the heat absorbed by the fresh, dry snowpack is much less than for wet snow or ice (Paterson, 1994), and melting is proportionately slower.

The need to develop a reliable proxy for change in a glacier's mass balance is crucial to paleoclimate reconstructions and future predictions based on glacial evidence. Previous studies have used the heat-budget model, calculating the difference between net radiation input and output in any given system such that leftover heat can be quantified (positive value corresponds to higher ablation rate) (Braithwaite, 1981). An accessible, reliable on-site data recorder capable of measuring various conditions such as air temperature, wind speed, surface temperature, cloud cover, net radiation, precipitation and moisture content is ideal for most accurately assessing conditions at the glacier (Naftz et al., 2002), but rarely have been deployed. In cases in which nearby data recorders do not exist or include only partial information, the statistical relationship between air temperature and glacier mass balance may instead be used to estimate either air temperature or glacier mass balance based on the known parameter.

Ice cores with an unaltered stratigraphy can potentially be useful in identifying mass-balance changes because they can record seasonal layers and the corresponding accumulation rates. Linking net mass-balance data from ice cores to the corresponding meteorological changes would provide a powerful tool for reconstructing the local climate history by proxy. Many factors can influence the mass balance of a glacier, but summer temperature and winter precipitation are the primary controls (Letréguilly, 1988), especially in the Sierra Nevada (Danskin, 1998). Letréguilly (1988), however, found that

the relationship between mass balance, summer temperature and winter precipitation for small mountain glaciers in Canada is highly variable, even locally between neighboring glaciers. This variability highlights the need to establish independently each study site's relationship to mass balance, winter precipitation and summer temperature before paleoclimatic reconstruction can be attempted. Once a mass-balance history is established, however, it is valid to extend that history to neighboring glaciers that experience similar effects (Letréguilly and Reynaud, 1990).

Braithwaite (1981) developed a simple model for estimating net ablation in glaciated areas for which little hydrological and meteorological data exist. This model relies on a statistically demonstrated relationship between air temperature and ablation, calculated from four glaciers in the Canadian Arctic. The largest input of energy for ablation is net radiation, greater than both sensible and latent heat. However, ablation correlates better with surface air temperature because of the variability of the sensible heat flux. This relationship is reasonable, given that sensible heat flux is proportional to air temperature. The average increase in ablation rate due to temperature is $6.3 \text{ mm day}^{-1} \text{ K}^{-1}$ (Paterson, 1980). Braithwaite (1981) developed the best current method of relating air temperature to annual ablation rates, using regression equations to correlate air temperature and ablation. Air temperature is assessed as the sum of positive degree days where

$$S = \sum_i \alpha_i T_i \quad (6)$$

T_i = mean daily temperature (K)

$$\alpha_i = 1 \text{ if } T_i \geq 0^\circ \text{C}$$

$$\alpha_i = 0 \text{ if } T_i < 0^\circ \text{C}$$

The ablation season typically lasts less than four months in the areas Braithwaite studied, and the temperature during the rest of the year is irrelevant. At best, this method can estimate about half the variance in ablation rate, but in cases where air temperature is the primary data available (such as the Sierra Nevada), it is the most reliable method of estimating ablation.

The stable isotope stratigraphy and temperature-mass balance correlations of Upper Fremont Glacier in the Wind River Range, Wyoming, have been extensively studied (Naftz et al., 2002). In order to link on-site air temperatures with $\delta^{18}\text{O}$ values, Naftz et al. developed transfer functions using a nearby SNOTEL site in order to reconstruct historical temperature changes over the past few centuries. To further refine the transfer functions, the study was repeated using an onsite ultrasonic snow-depth and an air temperature sensor. The transfer function developed from the on-site sensor ($r^2 = 0.71$, $p = 0.0179$) was superior to that developed from the SNOTEL site ($r^2=0.65$, $p\leq 0.0001$). Although the on-site monitoring sensor allowed more accurate transfer functions by better characterizations of localized effects (post-depositional snow removal and accumulation) and is the best solution when available, the transfer functions developed from the SNOTEL site were still useful in constraining air temperature changes.

II Methods

The Palisade Glacier was inspected with air photos and on foot to select an appropriate coring site. The ideal location is within the accumulation area to minimize

potential loss of records because of ablation. Locations directly below couloirs and areas of high rockfall were avoided to prevent exaggerated snow accumulation and drill damage. The selected site is at Lat 37.0972 N, Long 118.5084 W, UTM zone 365955E 4106720N (WGS84/NAD83), at 3936.9±1.4 m asl (Figure 1, 2).

Two types of samples were collected: snowpit samples to characterize depth changes and snow water equivalence within the seasonal snowpack in the accumulation zone, and ice-core samples to analyze stable-isotopic and trace-element signatures. The initial snowpit samples were collected in early July 2003 at the pre-selected coring site. The glacial surface was marked by a heavy accumulation of dust, and a transition from snow/firn to ice. The pit was excavated 2.6 m and bottomed on the glacial ice surface. Subsequent snow pits were sampled in early August 2003 (~1 m) and late June 2004 (1.6 m). No late-summer 2004 pit was sampled because the site lacked snow cover then. Snow water equivalent was measured in the June 2004 pit using a 750-cm³ “wedge cutter” and balance. A water sample from the meltwater outflow was taken near Sam Mack Meadow concurrent with the 2004 snow samples. Stable isotopes were measured in the July 2003 snow pit (δD) and the June 2004 snow pit $\delta^{18}O$ samples.

Two adjacent shallow (4- and 6-m) ice cores were drilled at the Palisade Glacier site in late August 2003 using a modified CRREL hand auger (Figure 3). Textural/visual characteristics were recorded on site and the ice cores sawed into consecutive samples and secured in whirlpaks. The samples were loaded into an insulated core box, packed with dry ice, and hauled off the mountain. Drilling the initial ice core on Palisade Glacier was difficult, and snow contamination of the core may have occurred at several points during the drilling when the core stuck within the auger barrel and coring was

interrupted. Extraction of the second core was much smoother, with more coherent sections of ice core removed. Therefore, I used the second core for analysis.

Trace-element concentrations were measured at Desert Research Institute in Reno, NV using a standard batch-analysis ICP-MS, and stable-isotope ratios for both the ice-core and snow-pit samples were measured at the University of Washington Quaternary Research Center Stable Isotope Laboratory using a Micromass Isoprime dual-inlet mass spectrometer and standard CO²-equilibration techniques. In cases in which stable-isotope ratios were measured for only δD (the summer 2003 snow pit), the data were converted to $\delta^{18}O$ using the relationship between the two isotopes derived from the Palisade ice core.

Data for both temperature and precipitation originated from the California Data Exchange Center (CDEC), which posts hydrologic data from the California Department of Water Resources. The SNOTEL data station is located in the north fork of Big Pine River (BGP), at Second Lake (37.1280°N, 118.4750°W, at 2970 m asl. This is the nearest site to the Palisade Glacier on the eastern crest of the Sierra Nevada. The data initially exist as hourly temperature measurements (°F) for the years spanning 1989 through 2004, which includes the time represented by the short ice core retrieved from Palisade Glacier. Snow water equivalent is recorded in inches. The SNOTEL data sensor error is calculated at $\pm 0.2\%$ at 20 °C (D. Hart, California Cooperative Snow Surveys Department of Water Resources, pers. comm., July 2005). From October 20, 2003 to September 10, 2004, a data logger was placed in a bedrock joint above the bergschrund on Palisade Glacier (J. Sanders, pers. comm., 2003) to gather hourly temperature readings. These temperature readings are used for comparison to the

SNOTEL data, although the 1-year record prevents its use for my study. Precipitation measurements were compiled as snow water equivalents (in inches) to minimize the effect of variable snow density.

The summer ablation season was defined by using CDEC snow water equivalent readings to identify the beginning and end of snow cover at the BGP site, as that is the closest data source to the glacier. The summer ablation season is therefore defined as beginning May 1 and ending Nov 1. This period was applied equally to each year for consistency. Mean daily temperatures are sufficient for positive degree day (PDD) calculations (Paterson, 1980, p.74), so the data were averaged to daily means and converted into °C for consistency. Periods with missing data were filled by interpolating between the endpoints of each gap, so that temperatures change in equal increments from one endpoint to the other. In most cases, only one or two mean daily averages are missing, and no more than seven occur in any series. This allows consistency in comparing yearly PDD.

Positive degree days were calculated by listing mean daily temperatures of each ablation season in an Excel spreadsheet and summing the values of days in which the mean daily average equals or exceeds 0.00 °C. In two years, 2001 and 2002, large portions of the snow accumulation record are missing. In order to include these years for PDD comparison, the relationship between the partial ablation season and the entire ablation season was determined as a ratio of partial to complete seasons based on years with complete data. On average, the PDD for the entire season was 1.3 times the PDD for the partial season. This allows assumed values of PDD for 2001 and 2002 based on the available data for those two years.

Ideally, PDD measurements would be from on-site measurements of air temperature. Although no such data exist for our coring site, a temporary data logger in a rock fracture above the bergschrund provides nearly one full year of temperature data (Lat 37.09 N, Long 118.508: J.Sanders, written comm., 2003) (Figure 4). Although this instrument was only in place less than a year and temperatures were likely strongly influenced by nearby rocks, there is still a reasonably good correlation between the two sets of temperature measurements ($r^2 = 0.76$).

The lapse rate, (change in temperature with elevation) is not established for this area of the Sierra Nevada. To constrain it, I compared temperatures above the Palisade bergschrund and the local SNOTEL site over nearly a year, and estimated monthly lapse rates calculated based on averaging differences in temperature for each month of the ablation season (Figure 5). Using these estimated monthly lapse rates, I estimated PDD values from temperature values at the Palisade Glacier coring site.

The relationship between PDD at the SNOTEL site and PDD at the Palisade Glacier is not reliably predictable using this method; the PDD at SNOTEL varies by a factor of 2.00 to 2.51 compared to that of Palisade Glacier (Figure 6). Additionally, direct comparison of the PDD at Palisade Glacier for 2004 (from the in-fracture data logger measurements) and those at the SNOTEL site does not support the lapse-rate method for the ablation season. The 2004 summer ablation season PDD at the coring site as predicted using the calculated lapse rates extrapolated from the SNOTEL site is 729°C; the PDD as indicated by the local in-fracture data logger for the same time period is 226°C, a difference of nearly 300%. Therefore, because of this substantial disagreement the following calculations and comparisons rely on the original data from the SNOTEL site, despite the fact it is 975 m lower than the study site. This method

reduces error because it relies on a normalized relationship between PDD at the SNOTEL site and conditions at Palisade Glacier.

III Results

Ice Core /Snow Pit Samples

The Palisade ice cores varied visually and texturally with depth. The ice core used for sample analysis was comprised of 0.05 m of clear ice, underlain by 0.12 m of wet, loose firn. Clear, uniformly bubbly ice dominates the rest of the core, except for another loose firn layer between 0.85 and 1.27 m depth and a horizon of cloudy, bubbly ice between 1.75 and 1.77 m depth. Two snow pits (excavated 7/10/03 and 8/6/03) preserve alternating layers of snow and ice lenses, bottoming in a prominent dust-rich bed at the snow/ice transition. The single snow pit dug the following summer (6/28/04) contained 0.02-m ice lenses at 0.10 and 0.52 m depth (total snowpit depth: 1.60m), with an abrupt transition to solid, bubbly ice at the bottom. In all cases, the snow was wet and dense, with a coarse texture. The average density of snow in the 2004 snowpit was 704 kg/m³, with values varying from 613 kg/m³ to 777kg/m³.

Stable Isotopes

The Palisade ice yielded two datasets from each ice core, stable-isotope ratios and trace-element concentrations. Both vary with depth in the core. Despite the proximity of the two cores, the stable-isotope profiles do not match well, indicating that at least one of them was disturbed during or since coring. Therefore, the second core was used for this study, although contamination is possible for this core as well. The second core is 5.64 m long. Six isotopic maxima (peaks) occur in the core, at 0.11, 1.16, 1.93, 3.05, 4.45, and

5.64 m depth (Figure 7). The five adjoining stratigraphic isotopic minima (troughs) occur at 0.95, 1.58, 2.88, 3.42, and 5.23 m depth. $\delta^{18}\text{O}$ values vary from -16.77 to -13.99 ‰, and δD values vary from -133.39 to -111.01 ‰ (33 measurements total). These values are similar to previous studies on snow in the Sierra Nevada, in which the $\delta^{18}\text{O}$ varies from -5 to -26 ‰ (Krouse and Smith, 1973). The standard deviation of the time-corrected sample standards is 0.99311 ‰. The typical difference between maxima and minima in the $\delta^{18}\text{O}$ stratigraphy is close to 1.1 ‰. Stratigraphic variations in $\delta^{18}\text{O}$ do not appear to match textural changes in the ice core noted above.

The $\delta^{18}\text{O}$ profile measured in the June 2004 snow pit is distinct from those of the previous summer. The $\delta^{18}\text{O}$ values from the 2003 summer snow pit vary from -13.47 to -15.81 ‰, with an average value of -14.49 ‰. $\delta^{18}\text{O}$ values from the 2004 pit vary from -14.98 to -20.21 ‰, with an average value of -17.98 ‰ (Figure 8). $\delta^{18}\text{O}$ measured in the outwash near Sam Mack Meadow was -18.47 ‰, which is well within the range of snowpack values from the Palisade Glacier that year.

Trace Elements

Concentrations of 18 elements (15 metals, two metalloids and one non-metal) were measured in sample splits from the stable isotope analyses. The concentrations vary with depth similarly to the stable isotopes. The average concentration profile of all the trace elements has six maxima at 0.025, 1.16, 1.93, 3.23, 4.23, and 5.42 m depth. Trace element concentration minima occur at 0.56, 1.58, 2.88, 3.42, and 4.45 m depth (Figure 9). All measured trace element concentrations co-vary with slightly different

magnitudes. As with the stable isotopes, stratigraphic variations in trace element concentrations do not follow textural changes in the ice core.

Temperature and Precipitation

The average accumulation season in the Big Pine drainage basin lasts from November 1 to April 30, according to snow depth measurements at the Second Lake BGP SNOTEL data recorder site (<http://www.cdec.water.ca.gov>). The average daily temperature during the ablation season (May 1 through October 31) at Second Lake is 3.7°C. Average minimum daily values vary from -15.7°C to -6.8°C, and average maximum daily temperatures vary between 10.5°C and 17.2°C. Peak annual temperatures occur consistently at the end of the first week in August. Minimum annual values average -11.6 °C, occurring anytime between December and March. Annual minima/maxima magnitudes do not appear to correlate from year to year ($r^2 = 0.06$, $p = 0.7$). The temperature data logger placed above Palisade Glacier (J. Sanders, pers. comm., 2003) recorded minimum, maximum, and average daily temperatures of -7.9°C, 6.4°C, and 0.17°C between May 1 and September 10, 2004 (Figure 4).

Annual winter snow accumulation varies significantly in the region near Palisade Glacier. Since 1989, the maximum snow accumulation recorded at the BGP SNOTEL site was 0.96 m s.w.e. in early April 1993. The lowest maximum snow accumulation was in 1994, when the snow water equivalent never exceeded 0.23 m. Large accumulations also occurred in late April 1995 (maximum value of 0.91 m), mid-April 1998 (0.70 m) and 0.70 m in mid-February 2005 (data were not recorded between 2/19/05-7/22/05) (Figure 10). The average snow accumulation between 1989 and 2004 is 0.44 m we. Annual snowpack begins to accumulate between mid-October to late November, with the

rate of accumulation increasing significantly by February, peaking in mid to late April, and melting rapidly in late April. Low-snow years follow the same pattern but the accumulation generally starts later and melting begins earlier.

IV Analyses

Stable Isotopes

The relationship between $\delta^{18}\text{O}$ and δD in the Palisade ice core is $\delta\text{D} = 7\delta^{18}\text{O} - 15$ with an r^2 of 0.95 (Figure 11). The trend resembles the Global Meteoric Water Line (GMWL) ($\delta\text{D} = 8\delta^{18}\text{O} + 10$). This similarity suggests that the stable isotope stratigraphy has not been substantially disturbed since deposition. I therefore focus my analysis on $\delta^{18}\text{O}$ to represent both $\delta^{18}\text{O}$ and δD .

Deuterium excess d (calculated as deviation from the GMWL by equation 5) in the Palisade ice core varies between -1.84 and 4.1 ‰, with typical variations between minima and maxima averaging ca. 2.5 ‰. The deuterium excess value can reflect several processes such as moisture sources (Dansgaard, 1964) or post-depositional modifications in the snowpack (Thompson et al., 1995). The pattern of deuterium excess resembles the stable-isotope profile (Figure 12), and differs little from the GMWL value of 10 ‰. Therefore, the variation in deuterium excess probably primarily reflects moisture sources rather than fractionation within the accumulated snowpack.

As previously noted, $\delta^{18}\text{O}$ measurements collected from seasonal snowpacks on the Palisade Glacier are distinctly different between 2003 and 2004 (Figure 8). The stable-isotope ratio value of the lower part of the early summer 2003 snowpack is similar

to the stable-isotope values at the top of the ice core. However, the 2004 snowpack has more negative and more variable $\delta^{18}\text{O}$ values than either the ice core or the 2003 snowpack. The maximum stable-isotope ratio value of the 2004 snowpack is within the range of maximum values for the ice core. Better constraints on the long-term isotope variability of snowpacks at the Palisade glacier would require a more intensive study. However, it seems reasonable that the 2003 and 2004 snowpacks would provide distinct annual layers within the glacier ice if they were preserved.

Trace Elements

Other than the surface dust layer, no other visible dust layers occurred within the Palisade Glacier ice cores. I therefore measured trace-element concentrations as a proxy for summer dust accumulation surfaces. This provides a tool to identify annual layers independently of stable isotope records. The most common element in rock dust, Al, was selected along with Sr, Ba, V, and Mn to represent dust input during the summer dry deposition period. Concentrations of these five trace elements resemble the stable-isotope concentration variations within the Palisade ice core most closely, yet come from different sources. Aluminum is a common dust component contributed from bedrock, strontium and barium are primarily added by ocean spray in the moisture source, and manganese and vanadium represent pollution sources (J. McConnell, pers. comm., 2004).

The concentrations of trace elements on summer accumulation surfaces should reflect the rate at which the elements were taken up in air (Dick and Peel, 1985). The

best method presently available to test whether this is true for Palisade Glacier is to compare the concentrations in the probable source areas to the concentrations in the ice. The major storm systems that bring precipitation to the Sierra Nevada are southwesterly and pick up dust from western California as the fronts move in from the Pacific Ocean. Although there should be a background global dust flux, the majority of material is probably regional in origin, particularly because the next-nearest source of terrigenous dust is across the Pacific Ocean. Trace-element concentrations of typical California soils were compared to those in the Palisade Glacier ice cores in terms of trace-element enrichment, defined by average crustal abundances and normalized to Al, the most common element.

$$\text{Therefore, } E_n = \frac{\left(\frac{El_{Sm}}{Al_{Sm}} \right)}{\left(\frac{El_{Cr}}{Al_{Cr}} \right)} \quad (7)$$

where E_n is the enrichment value of the sample (Sm) compared to average crustal (Cr) concentrations. Values ranging from 0.1 to 10 are considered normal while those below and above are depleted and enriched, respectively (J. McConnell, pers. communication, 2005). Soil trace element abundances are from the Kearney Foundation special report (1996); all soils from western California were used in the calculations.

All of the index elements have values within the normal range (Figure 13), and the additional element Pb is significantly enriched. The Pb enrichment value of 34 is typical for post-1985 snow accumulations (J. McConnell, pers. comm., 2004), which defines the upper age limit of the ice core. Maximum concentrations of trace elements

denote original annual summer snow ablation surfaces, and if unaltered, maxima in trace element concentrations should coincide with those in the stable isotope profile.

Meltwater infiltration or significant fractionation of the snowpack could disperse the trace elements differentially, and significantly obscure the relationship between annual dust accumulation surfaces and stable isotope maxima. In order to compare stratigraphic records from the Palisade ice core, the data were resampled using Matlab software so that data points from $\delta^{18}\text{O}$ values and trace element concentrations (ppb) occur at regular, concordant intervals with depth (every 2 cm). In order to identify potential vertical shifts in the profiles, the two records were systematically offset and correlated using Matlab (Figure 14). The highest r^2 values occur at offset values of 4 cm ($r^2 = 0.1307$, $p = 0.0297$) and -18 cm ($r^2 = 0.2843$, $p = 0.0000$). Positive lag values mean the isotope stratigraphy is shifted down relative to the trace element stratigraphy, and negative lag values shift the isotope stratigraphy relatively up. The most significant p value occurs when the r^2 value is highest, and the least significant p value occurs when r^2 is zero. Although the r^2 values themselves are not high, the variation with lag distance shown in Figure 14 appears to be meaningful.

Dust accumulation occurs year-round, but reaches peak concentration on the glacier surface during the summer months when precipitation is largely absent. The positive offset of 4 cm between isotope and trace element (dust) maxima is most likely to be correct because the stable isotope maxima occurs in the warm spring precipitation accumulation, prior to (stratigraphically below) the maximum concentration of dust on the glacier surface. Besides the vertical offset in the ice core, the pattern and spacing of trace element data maxima are similar to that of stable-isotope data maxima (Figure 9).

Six trace element concentration/stable isotope ratio maxima and five minima occur, as noted in the results section. Although not every annual snowpack survives the ablation season, the similarity of two types of environmental data suggests the surviving annual layers are reasonably well-preserved.

Temperature/Precipitation

The sum of annual positive degree days (PDD) from the SNOTEL site “BGP” at Second Lake (2970 m) vary from 632°C·days to 1001°C·days, with an average value of 838°C·days. The warmest years between 1988 and 2004 are, in order, 1994 (1001°C·days), 1989 (992°C·days), 2003 (979°C·days) and 2000 (930°C·days). The coolest years during the same time span are 1995 (632°C·days), 1993 (701°C·days), 1999 (744°C·days), and 1998 (746°C·days) (Figure 15). In general, the first average daily temperature below 0°C at the BGP SNOTEL site occurs by mid-September, and the average daily temperature rises above freezing by late May or early June.

The BGP SNOTEL site provides a record of winter precipitation in snow-water equivalent (we) for the study area. Annual snow-water equivalent is defined as precipitation occurring between Nov 1 to May 1 of the following year, and summed by daily average, such that the accumulation year of Nov 1988 to May 1989 is termed “1989.” The correlation between winter accumulation and temperature the following summer, while significant, is not strong enough to use as a predictive tool ($r^2= 0.45$) (Figure 16). The years with the most total snow accumulation were 1992/1993 (107.8 m.) and 1994/1995 (95.3 m), while the least were 1993/1994 (15.5 m), 1989/1990 (18.9 m) and 1991/1992 (19.7 m).

Local conditions dictate whether winter precipitation or summer temperature primarily affects glacier mass balance, which can vary within the same mountain range (Letréguilly, 1988). A more detailed study of on-site snowpack snow water equivalence and air temperature is needed to determine which factor controls mass balance on Palisade Glacier. However, it is reasonable to assume that years with large winter accumulation and cooler summer temperatures (such as 1992/1993) added to the mass balance, whereas years with reduced winter accumulation and warm summer temperatures (e.g., 1989/1990) lost mass. The years spanning 1993 to 1998 have relatively high accumulation rates and relatively cool summers; the years from 1989-1992 and 1999-2003 are warmer and had less winter accumulation. This is consistent with our ice-core records, which record five years of positive balance, likely spanning 1993-1998.

Snow water equivalent measurements of a snowpack on the Palisade Glacier collected concurrently with stable isotope samples in June 2004 indicated that the average density of the snow was 704 kg/m^3 . The snow was heavy and wet, and according to Paterson (1994, p. 9), wet snow and firm typically varies between $700\text{-}800 \text{ kg/m}^3$. With a snowpack depth of 1.4 m, this corresponds to 0.95 m we.

Although I did not measure it directly, I estimate the density of the clear, bubbly ice in the Palisade core at 850 kgm^{-3} (after Pohjola et al., 2005). The average thickness of annual layers in the ice core (as defined by trace element concentrations) is 1.07 m, equivalent to 0.9 m we. Therefore, the 2004 snowpack of 0.95 m we is equivalent to annual ice layers in the cores collected. However, snow accumulation during the winter of 2003/2004 was relatively low, and the timing of our sample (late June) was early in the

ablation season; in combination, these two factors suggest that little if any of the snowpack at the core site survived the 2004 ablation season.

The maximum accumulation during the 2003/2004 accumulation year was 0.37 m we (14.57 inches we) at the BGP SNOTEL site. Without direct measurements of snowpack at Palisade Glacier, the accumulation lapse rate between the SNOTEL site and the glacier is unknown. The net accumulation at the Palisade Glacier core site may also be influenced by factors such as wind scour and avalanche deposits. However, the accumulation record for 2004 provides an estimate of the minimum snowfall needed to preserve an annual layer at the Palisade Glacier core site. The winters with maximum snow accumulation greater than 0.38 m we (15 inches) are 1997/1998 (0.69 m we), 1996/1997 (0.46 m we), 1995/1996 (0.50 m we), 1994/1995 (0.91 m we), 1992/1993 (0.96 m we), and 1989/1990 (0.48 m we). The coherence of the isotope and trace element records combined with information about maximum snow depths and relatively cool temperatures indicate the core most likely represents five years within 1992 to 1998.

V Discussion

The coherence of the stable isotope, trace element, and physical records within the ice core indicate that the Palisade Glacier contains useful paleoclimate and paleoenvironmental records. However, it is crucial to establish the time represented by the ice core so that the records can be placed in context. The Palisade Glacier is clearly shrinking, especially in the later half of the 20th Century, reflecting substantial long-term negative mass balance; the climate causing this condition undoubtedly has caused repeated temporal hiatuses in the ice, even in the steady-state accumulation zone. Our

shallow ice core contains no clear stratigraphic marker horizons (such as ash layers), but a few time constraints can still be defined by comparing chemical stratigraphy and background pollutant levels to temperature and precipitation records. Lead levels in the ice indicate post-1985 ages for the entire core. The relatively shallow core depth and summer temperature/winter precipitation records indicate the five annual layers preserved are most likely from the mid-1990's. In support of this conclusion, the surface layer at the core site had a heavy dust layer on it; in contrast, there were no visible dust layer through the rest of the core, even where isotopic and elemental analyses indicated they exist. The fact that five low-snow years preceded our coring program suggests that the heavy surface dust layer actually represents five consecutive years with no net accumulation at the site, resulting in five years of dust accumulating at the surface of the ice. It is also important to note that meltwater from the net ablation years following the time span represented in the core did not appreciably alter the stable isotope or trace element records in the preserved ice.

Temperatures cannot be reliably reconstructed from stable isotope stratigraphy until more is known about the relationship between stable isotopes and storm events in the Sierra Nevada. Krouse and Smith (1973) noted the difficulty of developing simple correlations between weather conditions and stable-isotope ratios. Additionally, the stable-isotope record in a temperate glacier is likely disturbed to some degree by percolating meltwater, obscuring the original isotopic values. The coherence of the stable-isotope, trace-element and deuterium-excess records in the Palisade Glacier ice core, however, belies this process; despite five consecutive years of complete melting of the surface snow, distinct annual stratigraphy of these species was retained in the ice.

This observation suggests that diffusive processes related to meltwater percolation do not always erase stratigraphy in such temperate alpine glaciers, although the core lengths are insufficient to measure the scale length of isotopic homogenization, and therefore it is hard to assess the degree to which the amplitude of the layers may have been reduced.

In the absence of more robust local climate constraints, the most useful information provided by the isotopic and trace element stratigraphy in the Palisade ice core is their delineation of the annual accumulation layers. These layers provide an indication of mass balance of the glacier. Linking such a mass-balance record to detailed on-site temperature and precipitation records (similar to the weather stations used by D. Naftz, written comm., year) could supply a powerful proxy for reconstructing local paleoclimate history. Additionally, an on-site weather station would allow calculation of local lapse rates for both temperature and precipitation, as well as estimates of the effects of wind speed, humidity, and insolation on glacial mass balance. This information is necessary for extending the climate history recorded in the glacier past the 1988 installation date of the BGP SNOTEL site. Although the depth of the Palisade Glacier is undetermined, if the short ice core represents ca. 10 years of accumulation history, it is reasonable to estimate a hundred-meter core may contain at least several hundred years of ice.

Further study of the Palisade Glacier is warranted for several reasons. Very few high-altitude low-latitude ice-core records exist in the western U.S., and the Palisade Glacier is the largest, southernmost glacier in the Sierra Nevada. It is even possible that a more extensive ice-core record from the Palisade Glacier could be linked to other records such as lake sediments or tree rings, which could improve the breadth and accuracy of

local paleoclimatic reconstructions. Other ice-core possibilities include the neighboring Southfork Pass rock glacier, which originated as an alpine cirque glacier but is thickly mantled with rock debris. As previously shown at the Galena Creek Rock Glacier in northwestern Wyoming (Steig et al., 1998), rock glaciers can preserve stable-isotope and annual dust bands like bare-ice glaciers, but may contain much longer-lived records as a result of the shielding effects of the debris mantle. Test analyses of ice from the Southfork Pass rock glacier, collected in late summer 2003, were unsuccessful because of accidental contamination of the samples. However, its location near the Palisade Glacier suggests that the Southfork Pass rock glacier could provide supporting information for any overlapping sequences in the ice cores, as well as constraints on local sensitivities of these glaciers to changes in temperature and precipitation.

VI Conclusion

The Palisade Glacier contains surprisingly well-preserved environmental records within the upper 6 m of ice sampled for this study. Trace-element concentrations and stable-isotope ratios of $\delta^{18}\text{O}$ and δD vary with depth. Peaks in the trace-element concentrations denote annual summer surfaces, which closely correlate with peaks in stable-isotope stratigraphy. Analyses of temperature and precipitation information from a local SNOTEL site indicate the ice core contains five positive-mass balance years from the mid-1990's. Although the relationship between stable isotope values and temperature at the site is not well understood, the stable-isotope and trace-element stratigraphies contain mass-balance information in the form of annual-layer thicknesses. The preservation of these environmental records in the ice provides motivation for more comprehensive coring efforts on the Palisade Glacier in the future. Combined with more detailed on-site temperature, precipitation and snowpack information, a complete core could provide paleoclimate information related to the last century or more.

VII. Figures

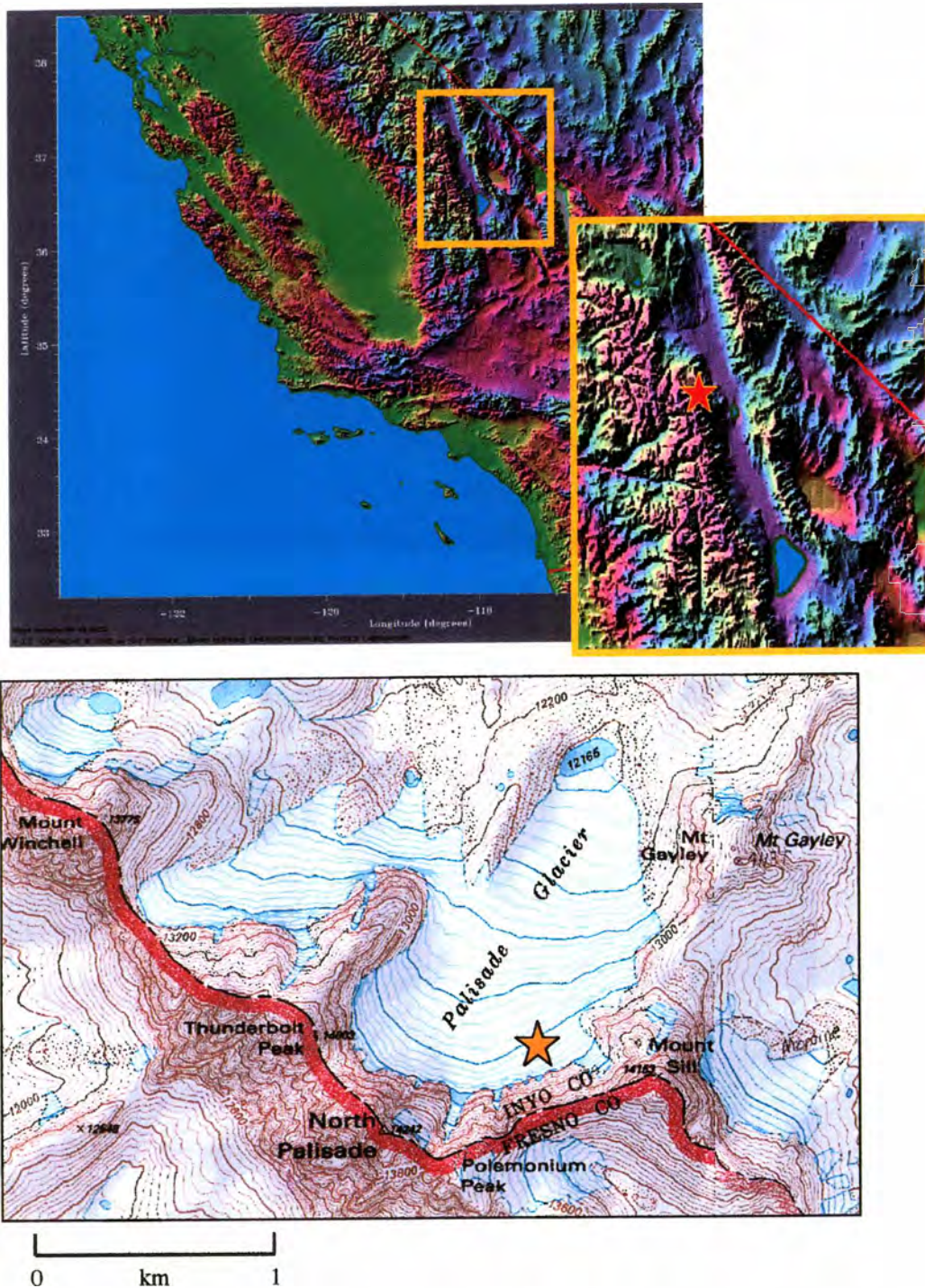


Figure 1
Palisade Glacier in the Sierra Nevada, CA. Star denotes coring location.



Figure 2

Photo from August 24, 1972. (Austin Post, U.S. Geological Survey Glaciology Branch, no. 72 R2-151). Yellow star denotes coring location. North is up. At the highest exposure of the moraines, the glacier is 700 m across.

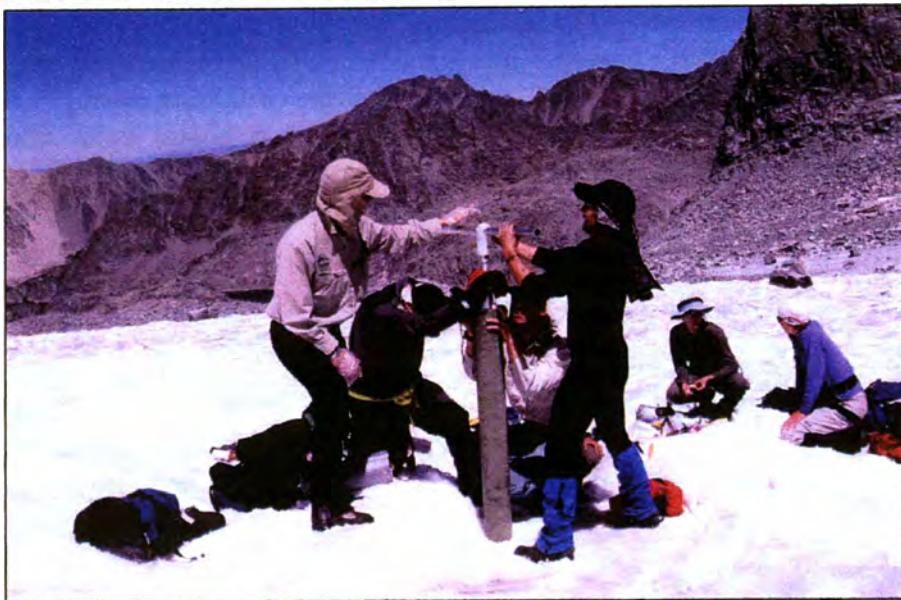


Figure 3

Coring the Palisade Glacier using a CRRL hand auger. View is to the south.

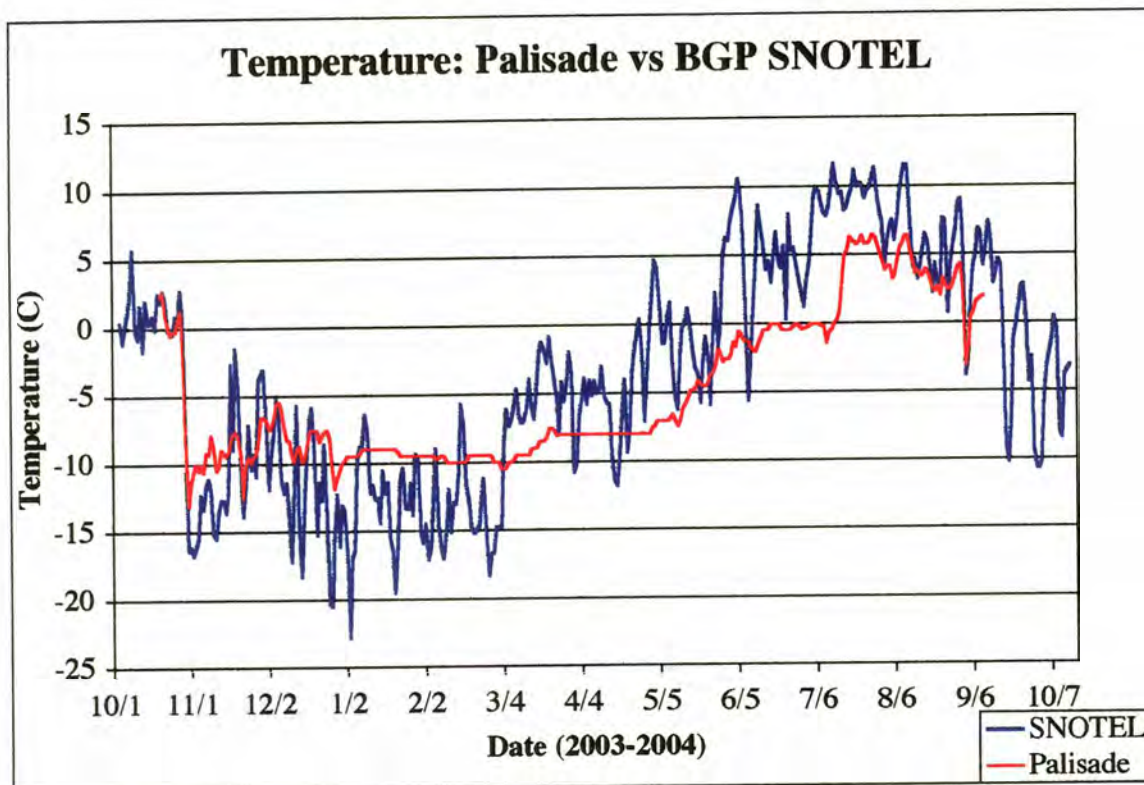


Figure 4

Relationship between SNOTEL (2970 m) and Palisade (3940 m) temperature readings from late September 2003 to early September 2004. The Palisade data logger recorded temperature from inside a crack in the bedrock above the bergschrund while the SNOTEL site recorded air temperature, which accounts for the decreased sensitivity of the Palisade data logger to air temperature.

Month	Lapse Rate, °C km ⁻¹
June	4.66
July	6.07
August	6.04
September	3.25
October	0.23

Table 1

Summer ablation season lapse rate values calculated between the Palisade data logger and the “BGP” SNOTEL site from 2004.

Year	PDD Ratio
1989	2.08
1990	2.02
1991	2.10
1993	2.49
1994	2.00
1995	2.40
1997	2.43
1998	2.03
1999	2.51
2000	2.26
2003	2.27

Table 2

Ratio between PDD values at the SNOTEL site and Palisade Glacier, using actual values at both sites for 2003 to predict values for Palisade Glacier based on the SNOTEL site. No data were available for 1992, 1996, and 2002. The average ratio value is 2.24 ± 0.24 .

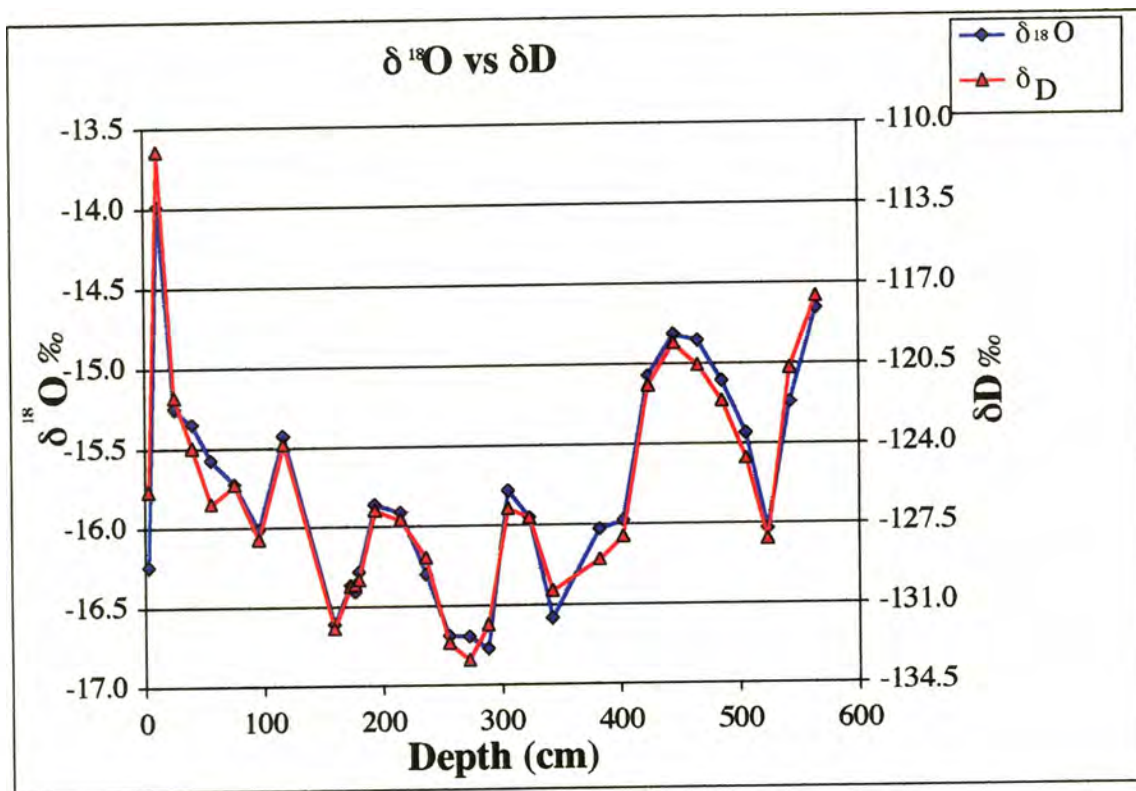


Figure 5

Stable isotope stratigraphy in the Palisade Glacier. The stratigraphy is similar enough between $\delta^{18}\text{O}$ and δD that following figures use the $\delta^{18}\text{O}$ stable isotope record to represent both.

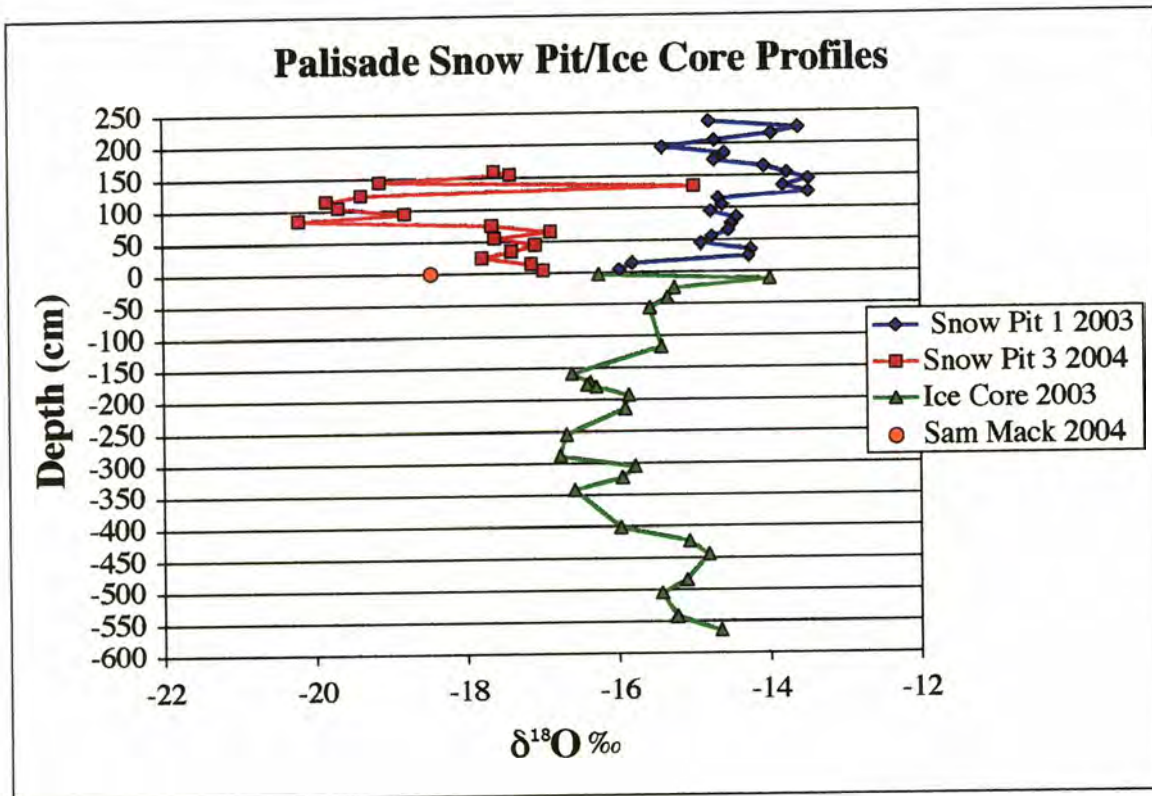


Figure 6

Palisade snowpit stable isotope profiles in comparison with the ice core. Positive Y values represent the snowpack; 0 is the base of the snowpack and the top of the ice core. A sample from the nearby outlet stream (Sam Mack) was collected at the same time as the 2004 snow pit and included for comparison.

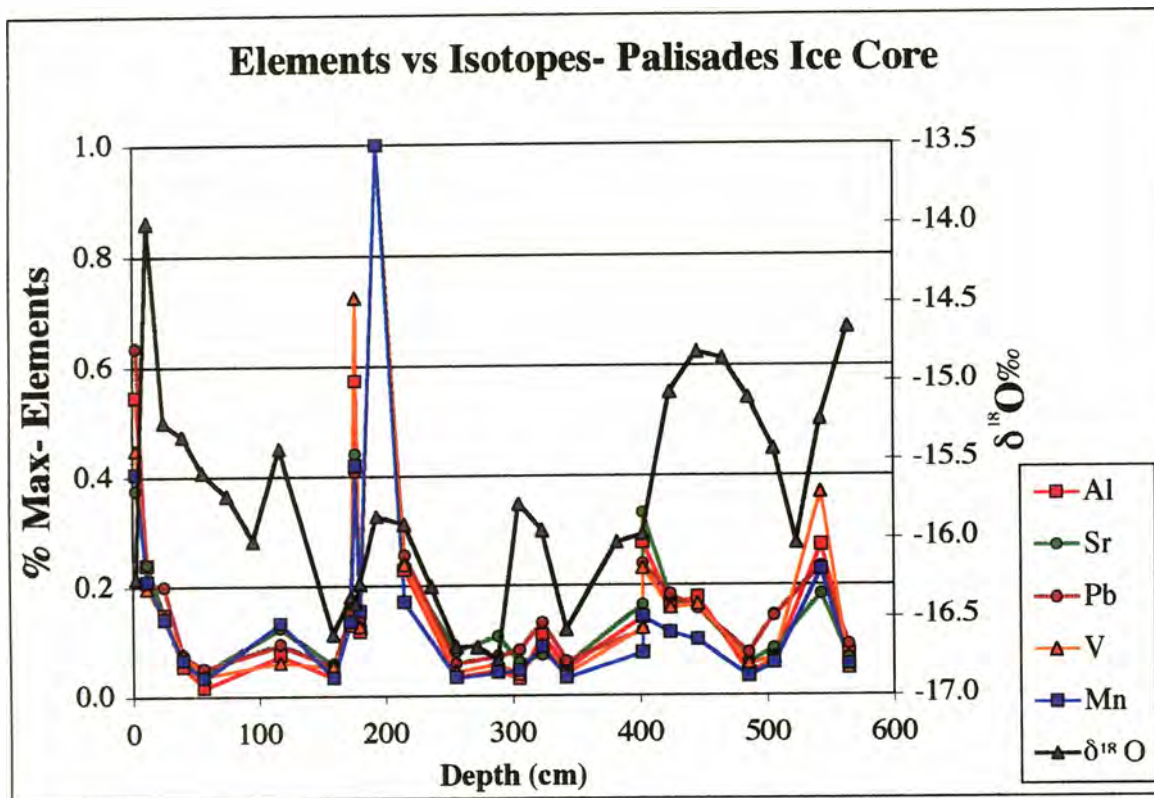


Figure 7

Trace-element (representative sample for clarity) stratigraphy versus stable-isotope stratigraphy in the Palisade ice core.

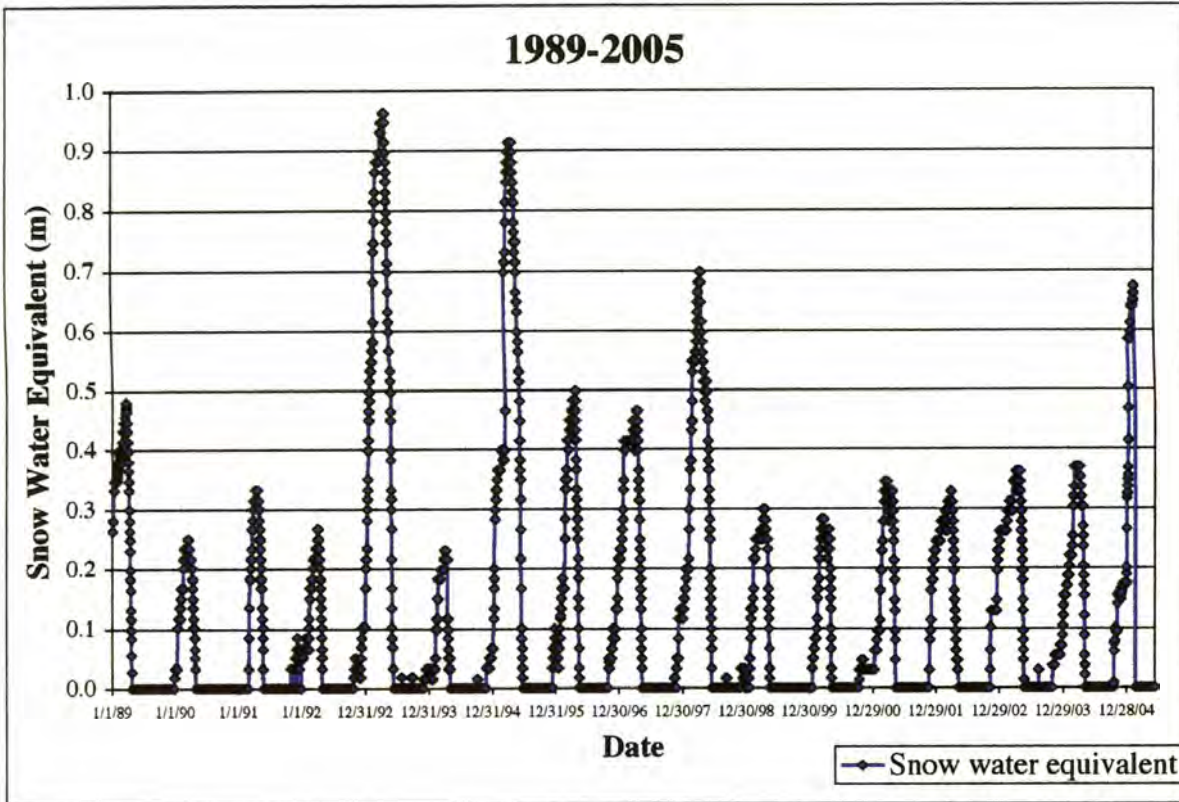


Figure 8

Snow water equivalent measurements from the “BGP” SNOTEL site, from 1989 to 2005.

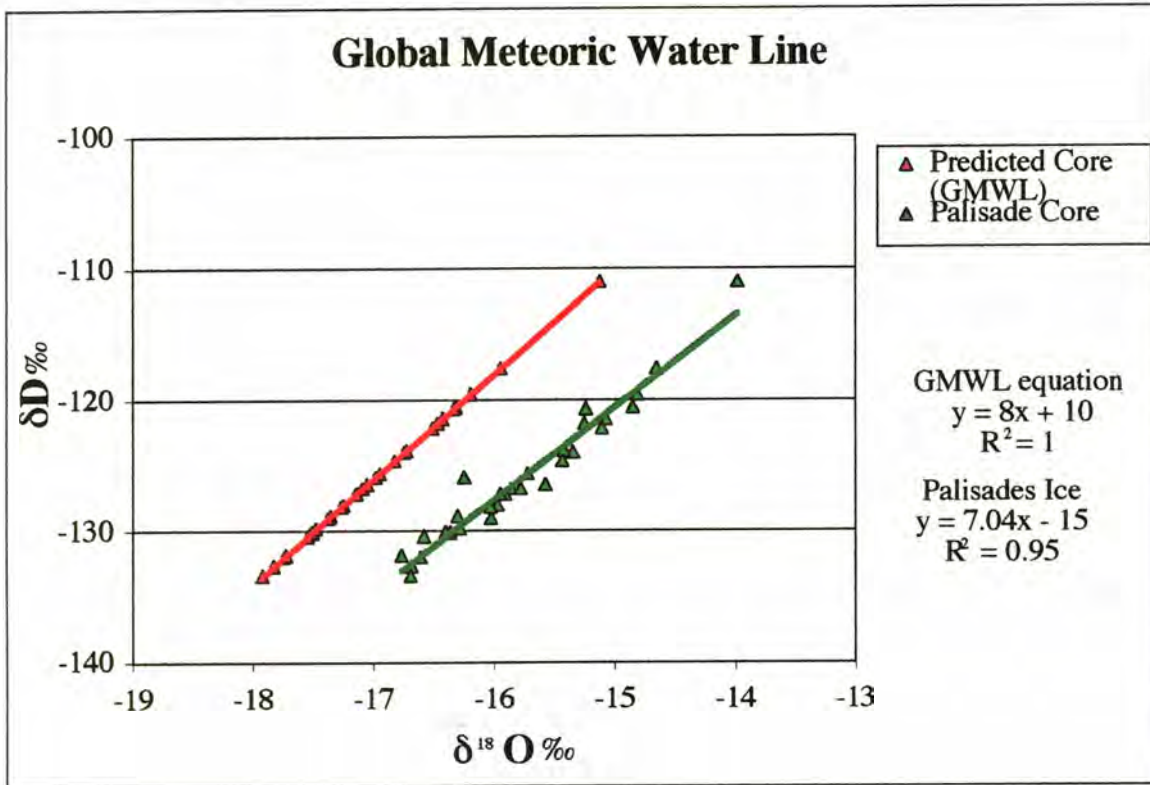


Figure 9

Relationship between $\delta^{18}\text{O}$ and δD in the Palisade ice core. The predicted relationship (red line) uses the Global Meteoric Water Line, and the actual relationship between $\delta^{18}\text{O}$ and δD is represented by the green line.

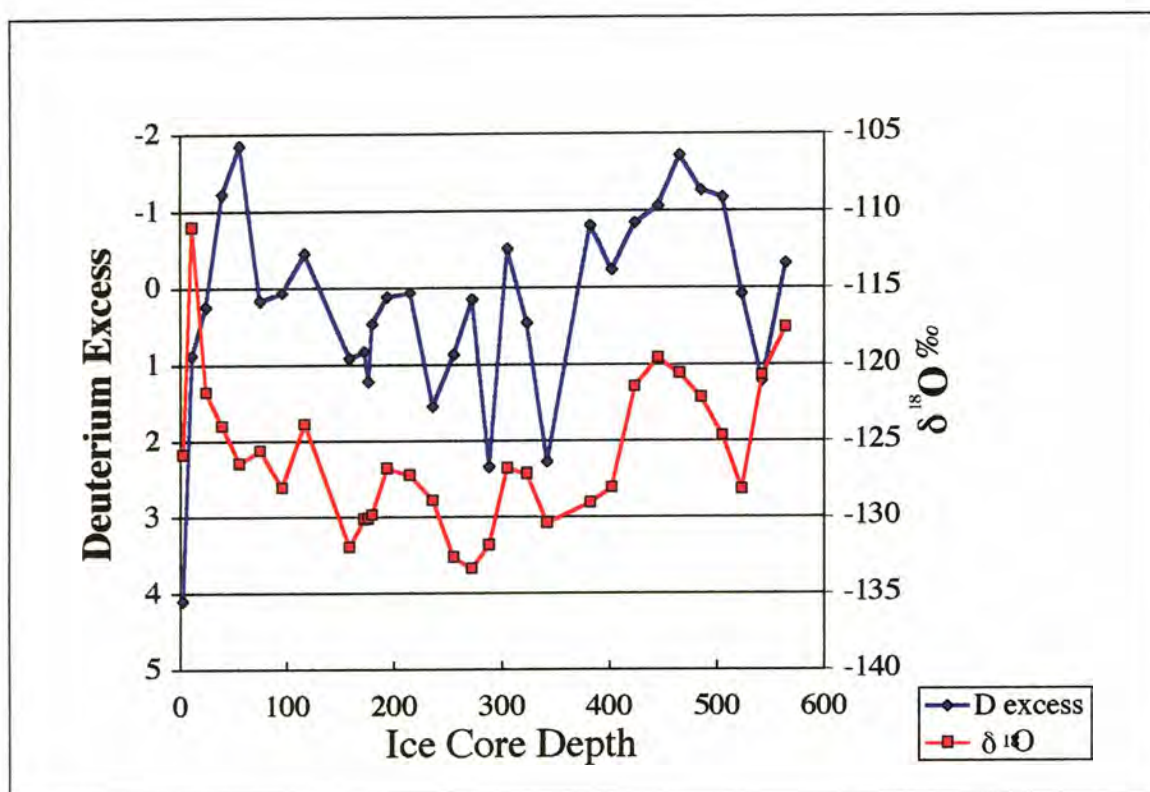


Figure 10

Deuterium excess (d) with depth in the Palisade Glacier and the relationship to stable-isotope stratigraphy.

	California Soils	Palisade
Element	Enrichment	Enrichment
Sr	4241	1.88
Ba	1.08	2.97
Pb	13701	34.7
V	20520	3.18
Mn	11899	3.78

Table 3

Representative trace element enrichment in Palisade glacier ice core.

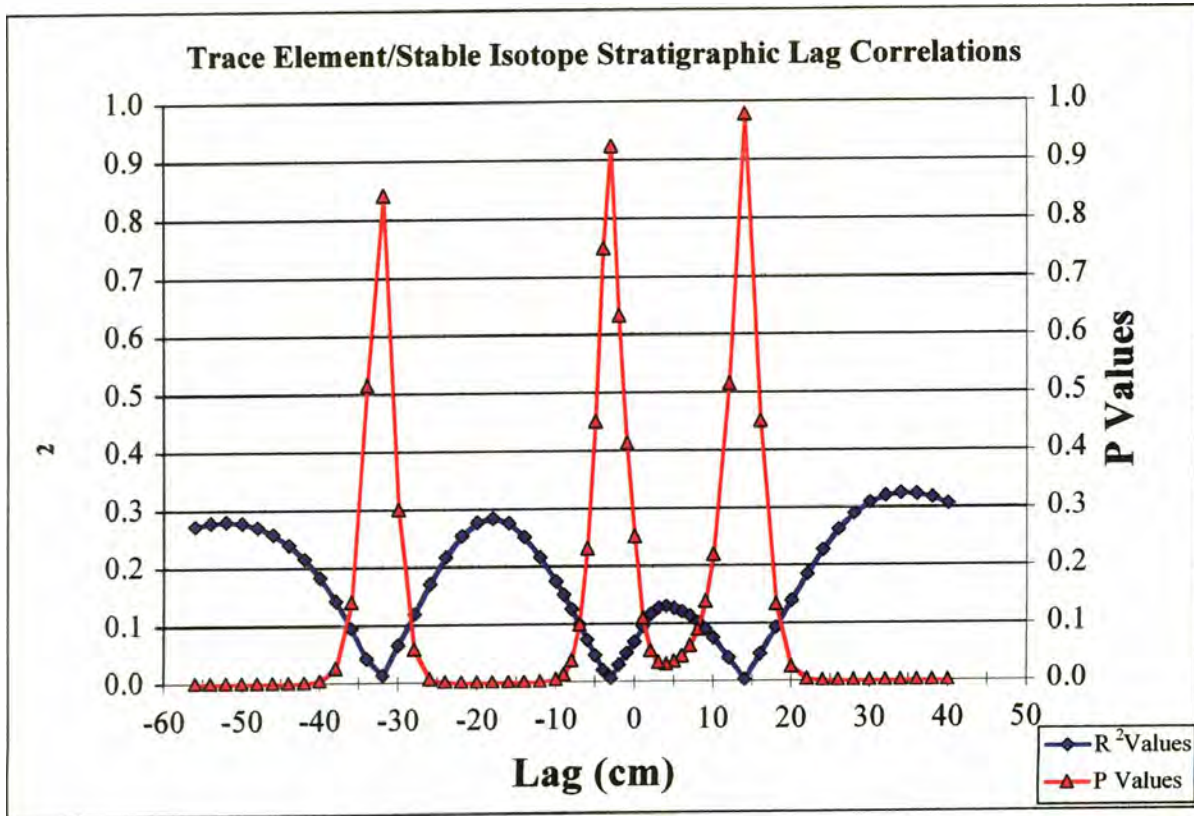


Figure 11

Matlab correlations of trace-element and stable-isotope stratigraphy for various offset amounts.

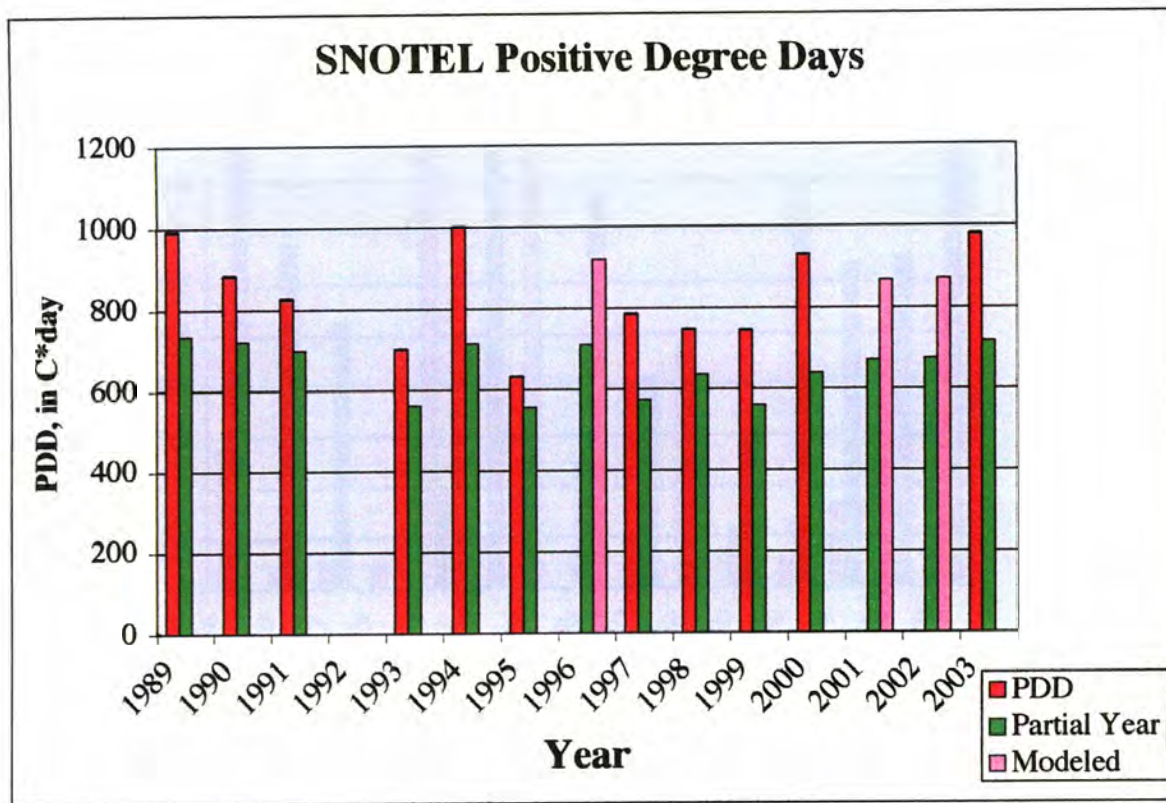


Figure 12

Positive degree days (PDD) from 1988 to 2003 at the “BGP” SNOTEL site. Pink bars are modeled years in which only partial data were available. No data existed for 1992.

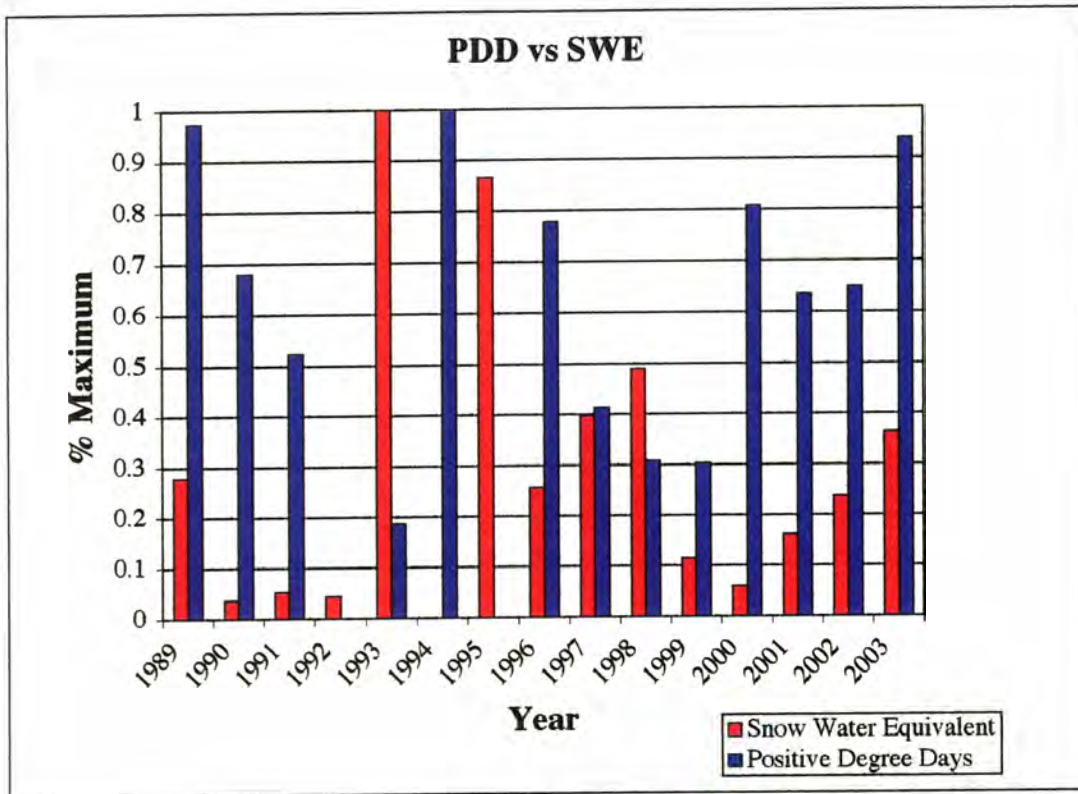


Figure 13

Comparison of winter precipitation (snow water equivalent) and temperature (PDD) for the following summer. Both parameters have been normalized to their maximum values (1989-2004).

VIII. References

- SNOTEL data from CDEC (California Data Exchange Center) at www.cdec.water.ca.gov
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