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EXTENT, TIMING, AND CLIMATIC SIGNIFICANCE OF LATEST PLEISTOCENE AND HOLOCENE GLACIATION IN THE SIERRA NEVADA, CALIFORNIA

by

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Abstract

Extent, Timing, and Climatic Significance of Latest Pleistocene and Holocene Glaciation in the Sierra Nevada, California

by Douglas Howe Clark

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Despite more than a century of study, scant attention has been paid to the glacial record in the northern end of the Sierra Nevada, and to the smaller moraines deposited after the retreat of the Tioga (last glacial maximum) glaciers. Reconnaissance mapping indicates that several previously undocumented Tioga ice fields and valley glaciers covered sizable parts of the Sierra Nevada north of the Tuolumne/ Stanislaus ice field. Equilibrium-line altitude (ELA) estimates of the ice fields indicate that the Tioga ELA gradients there are consistent with similar estimates for the southern half of the range, and with an intensification of the modern temperature/precipitation pattern in the region.

In the southern part of the range, high, north-facing cirques typically contain two distinct ages of moraines associated with the Recess Peak and Matthes (Little Ice Age) advances (Birman; 1964). Many Matthes moraines are voluminous and some form active valley-floor rock glaciers. Direct observations of ice exposed in these landforms, and theoretical considerations, indicate that most are cored by glacier ice, and that the surficial debris is a relatively thin (~1-10 m), continuous mantle that insulates the underlying ice from ablation. The rock glaciers are thus a form of debris-covered glacier; accumulation-area or terminus-to-headwall altitude ratios of such debris-covered glaciers, used to estimate ELAs, are substantially different than those of "clean" glaciers. Many Recess Peak glaciers, and a few Tioga glaciers, in the southern Sierra Nevada were apparently covered by debris. If the effect of debris cover is not recognized, ELA estimates for such former glaciers will be systematically too low, potentially by hundreds of meters.

The Recess Peak advance has traditionally been considered to be mid-Neoglacial age, about 2-3000 yr B.P., on the basis of relative weathering estimates. Sediment cores of lakes dammed behind moraines correlative with Recess Peak in four widely spaced sites yields a series of high-resolution AMS radiocarbon dates which demonstrate that Recess Peak glaciers retreated before ~13,100 cal yr B.P. (11,190 ± 70 ^{14}C yr B.P.). This minimum limiting age indicates that the advance predates the North Atlantic Younger

Dryas cooling. It also implies that there have been no advances larger than the Matthes in the roughly 12,000 year interval between it and the Recess Peak advance. This finding casts doubt on several recent studies that claim Younger Dryas glacier advances in western North America. The 13,100 cal yr B.P. date is also a minimum age for deglaciation of the sample sites used to calibrate the *in situ* production rates of cosmogenic ^{10}Be and ^{26}Al. The discrepancy between this age and the 11,000 cal yr B.P. exposure age assumed in the original calibration (Nishiizumi *et al.*, 1989) introduces a large (>19%) potential error in late-Pleistocene exposure ages calculated using these production rates.

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DEDICATION

To the memory of Clyde Wahrhaftig, who's keen insights into the glacial and geomorphic evolution of the Sierra Nevada and other alpine regions have guided and inspired numerous generations of geologists.

"Mountains are the culminating points of the scenic grandeur and beauty of the earth. They are so, because they are also the culminating points of all geological agencies."

Joseph LeConte, On Some Ancient Glaciers of the Sierra Nevada (1875, p. 128).

INTRODUCTION

Interest in past climate change has intensified during the past decade, in large part due to a host of new high-resolution proxy records that indicate Earth's climate systems may be less stable than previously thought (e.g., Shackleton and Opdyke, 1973; Alley *et al.*, 1993). In particular, long-term, continuous paleoclimate records measured in high-latitude ice cores and ocean sediment cores indicate that general circulation and temperature patterns shifted frequently between radically different states during the last glaciation, and that these fluctuations occurred over short time spans (i.e., decades or less)(Alley *et al.*, 1993). Potential mechanisms for these abrupt climatic shifts are still hotly debated (e.g, Hughes, 1992; MacAyeal, 1993; Denton and Hendy, 1994; Clark and Bartlein, 1995;). However, discerning viable mechanisms depends on whether the climatic shifts are local, regional, or global in extent. Because the ice and ocean sediment cores dominantly record regional temperature and global ice-volume conditions, other records that preserve local paleoclimatic information are needed to address the question of regional climatic shifts.

Alpine glacial landforms and associated lacustrine deposits offer two particularly useful sources of paleoclimate data. Mass balances of temperate alpine glaciers are highly sensitive to shifts in climate. Variations in seasonal distribution and amounts of precipitation, summer temperature, and cloudiness cause changes in the position of the average glacial snowline, or equilibrium line, the boundary between the zones of net accumulation and net ablation on a glacier. Under steady-state conditions, the annual mass balance at the equilibrium line is zero. By estimating equilibrium-line altitudes (ELAs) of former glaciers for different glaciations across a broad region, it is possible to detect both local and regional changes in integrated climate during successive glacial advances. Topography of former glaciers can be reconstructed using trimlines, moraines, and erratics, from which ELAs can be estimated, thus providing relative measures of climate (primarily temperature and precipitation) during different glacial events. When compared with modern conditions, it is possible to estimate absolute changes in these parameters (e.g., Leonard, 1989). Ice accumulation and ablation, the two primary mechanisms affecting ELAs, depend on climatic conditions within the relatively limited area occupied by the glacier and its adjoining drainage basin. These climatic conditions in turn relate to regional climatic patterns. By combining, from many valleys, the ELAs of a given period of glacier advance, one can reproduce regional ELA gradients or surfaces for a given climatic episode; comparing ELA gradients for different glaciations enables one to study potential changes in regional climatic conditions, including moisture sources, during different events. The primary limitations of ELA studies are: (1) they represent climatic conditions that existed during a relatively brief period of time when glaciers were at a maximum position; and, (2) the timing of such advances are notoriously difficult to date directly.

One method of overcoming these limitations is to study lacustrine deposits that can be related geomorphically to the glacial deposits, primarily in lakes that are dammed behind moraines or lakes that contain outwash from moraines formed higher in the drainage. Cores of such deposits nicely complement the "instantaneous" glacial record by providing continuous pre- and/or postglacial proxy records of climate change, such as sedimentation, vegetation (through macrofossils and palynology), lake conditions, and weathering rates. With the advent of AMS radiocarbon dating, lake cores provide a well-tested means for dating late-Pleistocene and Holocene sequences. Radiocarbon dates from cores of these lakes also supply limiting ages for adjacent moraines that cannot be dated directly. The largest drawbacks to these records for studying glacial records are that: (1) they are difficult and time-consuming to obtain, especially from high-altitude sites; and, (2) they typically provide more information about non-glacial than glacial periods, especially for palynologic and dating control. When used in conjunction with a thorough study of the glacial deposits and former ELAs, however, the lake records may help to decipher the glacial and climatic history of a region.

Glacial Climate Studies in the Western U.S.

Recent versions of general circulation models (GCMs) suggest that the climate of the western U.S. was wetter and colder than present during the culmination of the most recent glaciation (marine oxygen-isotope stage 2; e.g., CLIMAP Project Members, 1981; Kutzbach and Wright, 1985). This change resulted from a split, intensification, and southward migration of the north Pacific storm track from its current mean position near the California-Oregon border (CLIMAP Project Members, 1981; Kutzbach and Wright, 1985). Such GCM results are supported by recent paleoclimate studies of pluvial and vegetation records in the southwestern U.S. (e.g., Smith and Street-Perrott, 1983; Spaulding *et al.*, 1983; Smith, 1984).

Alpine glacial records in the mountains of western North America lend further credence to this theory (e.g., Porter *et al.*, 1983; Gillespie, 1991). However, recent ELA studies in the U.S. have either concentrated on interior mountains near the continental divide (Meierding, 1982; Leonard, 1984; Zielinski and McCoy, 1987; Locke, 1990), or have covered large regions by studying snowline trends that are integrated over many glaciations (e.g., cirque-floor altitudes; Porter *et al.*, 1983). Detailed ELA studies are sparse in the mountains of California, Oregon, and Nevada. Major mountain ranges of California, including the Sierra Nevada, the Warner Mountains, San Bernardino

Mountains, the northern Coast Ranges, the Trinity Alps, and the southern Cascades, preserve several sequences of glacial deposits and thus contain important information about changes in climate patterns in the western U.S. during the latest Pleistocene glacial maximum as well as during several earlier glaciations and subsequent Holocene advances.

The Sierra Nevada

The Sierra Nevada offers an excellent location to investigate glacial climate records at temperate latitudes and on a regional scale. It has a well preserved late-Pleistocene and Holocene moraine sequence that spans a wide latitudinal belt (ca. 6°). Moreover, its orthogonal intersection and occlusion of the modern Pacific winter storm track (Fig. 0.1) makes glaciers in the range particularly sensitive to the intensifications and southward migrations of the storm track predicted by recent general circulation models (CLIMAP Project Members, 1981; Kutzbach and Wright, 1985). Also, the proximity of the Sierra Nevada to the coast and its position as the first major mountain range encountered by such storms reduce the large-scale complicating effects of orography on climate. On the other hand, the range is at the western margin of the Great Basin. Because the southern Great Basin currently is subject to monsoonal (summer) incursions of moisture, glaciers in the Sierra Nevada may also react to substantial northward migrations of monsoons during periods of maximum insolation at temperate latitudes (Gillespie, 1991). Furthermore, the relatively simple form of the Sierra Nevada (generally a single crest) minimizes local orographic influences on the regional climatic snowline. A final factor is the abundance of tarns and moraine-dammed lakes in the range; these lakes provide a myriad of potential coring sites that are tied to the late-Quaternary glacial evolution.

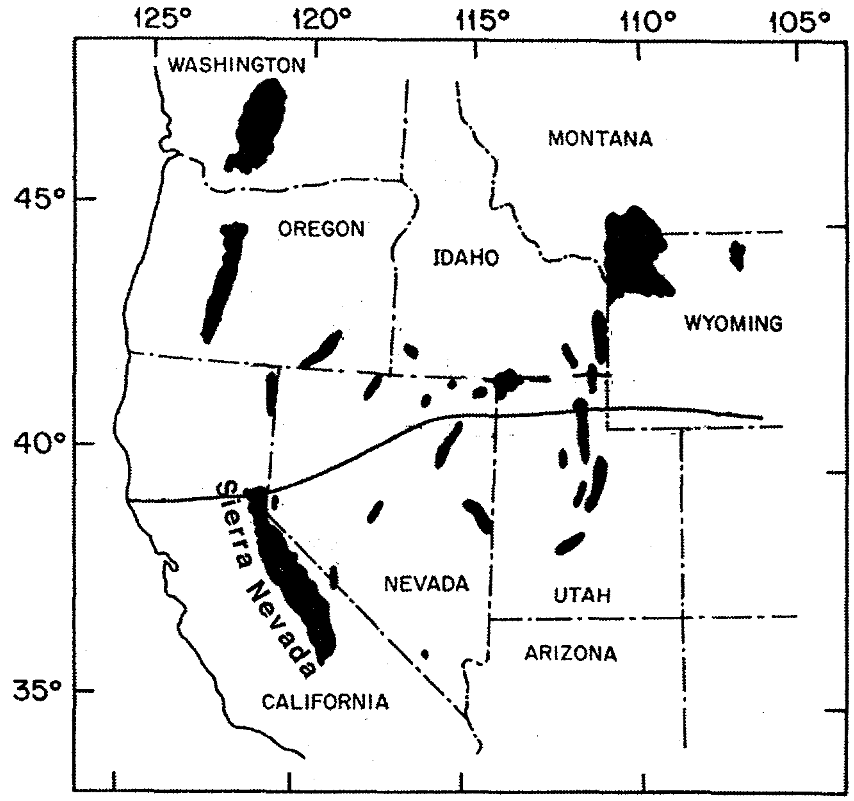


Fig. 0.1: Modern winter climatic boundaryies of western U.S. Dark areas represent major alpine regions. Solid line indicates January surface-wind convergence zone separating westerly winds from the Pacific north of the line from southerly and southwesterly winds south of the line (Mitchell, 1976). Areas north of the zone experience frequent intrusion of Pacific air masses during the winter; areas south of the zone are much less affected by westerly Pacific storm fronts. This boundary probably migrated several degrees of latitude southward during late-Pleistocene glaciation. Modified from Zielinski and McCoy (1987).

Glacial deposits in the Sierra Nevada, especially in the eastern and southern parts, have been studied for more than a century (e.g., LeConte, 1873; Muir, 1873; Russell, 1885; Gilbert, 1904; Matthes, 1930; Blackwelder, 1931; Sharp and Birman, 1963; Birkeland, 1964; Birman, 1964; Wahrhaftig and Birman, 1965; Clark, 1967; Sharp, 1968, 1969, 1972; Burke and Birkeland, 1979, 1983; Gillespie, 1982; Berry, 1994; Burbank, 1991; Gillespie, 1991). These studies suggest that the Sierra Nevada may have experienced up to six or more Pleistocene "glaciations" and up to three Holocene advances (compiled in a review article by Fullerton, 1986). The four youngest Pleistocene tills, Mono Basin, Tahoe, Tenaya, and Tioga (decreasing age), apparently represent regional late-middle Pleistocene through late-Pleistocene glacier advances, although uncertainties about numerical ages have led to persisting controversies over the exact timing and correlation of these events (e.g., Burke and Birkeland, 1979; see also section on Nomenclature below). Post-Tioga deposits have typically been subdivided into three Holocene (and/or latest Pleistocene) glacier advances: the Hilgard, Recess Peak, and Matthes (decreasing age; Birman, 1964; Yount *et al.*, 1982; Burke and Birkeland, 1983).

Despite the long record of investigation, several significant and intriguing gaps in our understanding of the glacial sequence of the Sierra Nevada remain. Most past studies focused primarily on local records, with only a few that attempted to investigate the climatic implications of the deposits on a larger, particularly range-wide scale (e.g, Gillespie, 1991). Furthermore, the bulk of the studies concentrated on characterizing the moraines formed during maximum advances; glacial deposits associated with the most recent deglaciation and subsequent cirque advances have remained relatively obscure. Similarly, glacial deposits and landforms in the northern end of the range near to and north of Lake Tahoe have received relatively scant attention.

Over the past several years, I have attempted to improve our understanding of the Sierran glacial record regarding some of these problematic topics. In this dissertation, I present and discuss my findings, which include: (1) reconnaissance field mapping and climatic interpretation of late-Pleistocene glacial deposits in the northern Sierra Nevada, north of about 38° N latitude (Fig. 0.1); (2) mapping and characterizing the climatic significance of post-Tioga debris-covered glacial deposits along the southern Sierra Nevada crest; (3) radiocarbon dating of lake sediment cores in conjunction with detailed glacial-geologic mapping to establish a more accurate chronology for post-Tioga advances in the Sierra Nevada; and, (4) synthesizing these data and those of other studies to support a new model for the timing and style of deglaciation of the range during the close of the Pleistocene. This lattermost point has important consequences for the current production-rate calibration of several cosmogenic isotopic systems that are based on an assumed rather than measured age of deglaciation of the Sierra Nevada (Nishiizumi *et al.*, 1989). Such isotopic measurements are rapidly gaining acceptance as surface-exposure dating tools in climatic and geomorphic studies.

Each chapter of the dissertation summarizes the results of a distinct portion of the study, and therefore each includes an abstract, introduction, and conclusion. The first chapter describes reconnaissance mapping of glacial deposits in the northern, and least-mapped, end of the range. During this study, I found strong evidence for several extensive late-Wisconsin (Tioga-equivalent) ice fields and ice caps and many valley glaciers that previously were poorly documented or unrecognized. ELA estimates from this mapping match well with and extend those compiled by Gillespie (1991) for the southern half of the range. Evidence for earlier glaciations also exists, but is more subtle and discontinuous than in the central and southern Sierra Nevada. The second chapter introduces cirque deposits in the southern end of the range and investigates the effects that the extensive debris cover on many such glaciers can have on reconstructed ELAs.

Similar debris-covered glaciers, although not abundant, probably existed locally in the Sierra Nevada during maximum glaciations as well. The third chapter presents the results of a coring study of lake sediments associated with moraines of the Recess Peak advance. The limiting radiocarbon dates from this portion of the study confirm that the advance was a late-glacial event and not a Neoglacial event as generally thought. Combined with regional mapping of cirque moraines, these findings indicate that the Matthes (Little Ice Age) advance, which culminated in the last 700 yr (Wood, 1977), was the largest of the Neoglacial period. The fourth and final chapter combines the radiocarbon dating control for the Recess Peak advance and that from other studies with geologic evidence for stagnation and early deglaciation of the Tioga glaciers to indicate that current late-Pleistocene/Holocene production-rate calibrations for some cosmogenic exposure-age dating methods (Nishiizumi *et al.*, 1989) may be in error by 20% or more. Such an error is supported by recent exposure age measurements on Laurentide terminal moraines in New Jersey, and by calculations of the effects of past paleomagnetic field strength variations on *in situ* cosmogenic production rates (Clark *et al.*, 1995 (in press)). These findings have distinct implications for studies that use cosmogenic isotope exposure ages, since they may nullify the conclusions of several recent studies that purport to correlate local moraine sequences to brief, regional climatic fluctuations, such as the North Atlantic Heinrich and Younger Dryas cooling events (e.g., Phillips *et al.*, 1990; Phillips and Zreda, 1992; Easterbrook, 1994; Evenson *et al.*, 1994; Clark and Bartlein, 1995; Gosse *et al.*, 1995a, 1995b). Most of the work presented in this thesis has already been published (Clark *et al.*, 1994a, 1994b; Clark *et al.*, 1995 (in press); Clark and Gillespie, 1995 (in press)).

Nomenclature

In spite of, or perhaps because of, the relatively long history of glacial investigation, the taxonomy of glacial deposits in the Sierra Nevada remains confusing. Because my work requires a reshuffling of the late-Pleistocene/Holocene glacial sequence in the Sierra Nevada, I attempt to clarify my use of terminology in this dissertation.

Drifts of the major Pleistocene advances in the Sierra Nevada have never been formally defined, with type areas and distinct stratigraphic and chronologic designations. In particular, it is difficult to relate the most recent deposits (Tioga, Tenaya, Tahoe; Sharp and Birman, 1963) to the global standard for glaciations that has been established with the marine oxygen isotope curve (e.g., Shackleton and Opdike, 1973) and to the standard North American glacial stratigraphy (e.g., Fullerton, 1986). To a large degree, the situation reflects uncertainties in the ages of various drift units. However, there is enough numerical age control to show distinctly that the Tioga, Tenaya, and at least some Tahoe tills were deposited during the most recent (Wisconsin) glacial cycle, which encompasses marine oxygen isotope stages 4, 3, and 2 (e.g., Gillespie, 1982; Phillips *et al.*, 1990; Gillespie and Molnar, 1995). Yet in the informal Sierran glacial designations, each of these advances is typically referred to as a "glaciation."

Thus there is no stratigraphic designation equivalent to the Wisconsin Stage in the Sierra Nevada, and it is unclear how Sierran "glaciations" relate to the more global stages. Furthermore, as other workers have noted, the Tioga/Tenaya/"late" Tahoe sequence forms a complex of very similar deposits (e.g., Blackwelder, 1931, p. 881; Porter, 1971, p. 317; Gillespie and Molnar, 1995, p. 318), which to some have been indistinguishable (Burke and Birkeland, 1979; Gillespie and Molnar, 1995, p. 319).

There is a fairly direct change that would improve the Sierran glacial nomenclature: define a local equivalent to the Wisconsin Glaciation. Thus, the late Tahoe, the Tenaya, the Tioga (both early and late), and, of importance to this study, the

Recess Peak advances would become stades of the Wisconsin-equivalent glaciation. This solves the problems created by the Recess Peak, which is too small to be called a "glaciation," and the Tenaya, which may actually be an early Tioga or late "late" Tahoe advance (Burke and Birkeland, 1979; Bursik and Gillespie, 1993). There are several type-areas/names for such a glaciation. The deposits in Mono Basin (e.g., the "Walker Creek Glaciation") are one such possibility. However, it is beyond the scope of this dissertation to decide this matter. The main import for the following discussion is that the unnamed Wisconsin-equivalent glaciation in the Sierra Nevada provides a framework in which to place the late-Pleistocene deposits. I therefore refer to the Tioga and Recess Peak advances as "stades" of the last (unnamed) glaciation. Although I don't directly deal with them, the "late" Tahoe and Tenaya advances would also be stades of the last glaciation. Moreover, both Tioga and Recess Peak stades appear to have occurred during late-Wisconsin time (Bursik and Gillespie, 1993; Clark and Gillespie, 1995 (in press)). Where discussing previous work, however, I will maintain the terminology used by the author(s).

CHAPTER 1: EXTENT OF LATE-PLEISTOCENE GLACIATION IN THE NORTHERN SIERRA NEVADA

INTRODUCTION TO CHAPTER 1

Late-Pleistocene glacierization in the Sierra Nevada has been described primarily along the eastern slope and in the southern half of the range, with the exception of the ice field that filled the Yosemite and Merced drainages (e.g., Matthes, 1930; Blackwelder, 1931; Matthes, 1960; Sharp and Birman, 1963; Birman, 1964; Clark, 1967; Sharp, 1969; Sharp, 1972; Burke and Birkeland, 1979; Gillespie, 1982; Burbank, 1991; Gillespie, 1991; Bursik and Gillespie, 1993). Near the southern limit of glaciation in the range, where the mountains reach their maximum altitudes, late-Pleistocene glaciers within the San Joaquin and Kern drainages were confined by deep canyons and overflowed divides only at small passes (Matthes, 1965; Janda, 1966). Several studies have documented the large ice field that dominated the crest of the range at the headwaters of the Tuolumne and Merced drainages during late-Pleistocene glacial ages and have noted that it was substantially transfluent to the east side of the range (Russell, 1889; Matthes, 1930; Wahrhaftig and Birman, 1965; Clark, 1967; Sharp, 1972).

It has generally been assumed that the Tuolumne ice field was the largest transfluent ice field in the range, where the southward increase in altitude balanced with the southward decrease in precipitation to produce the maximum glacier size (Matthes, 1965; Wahrhaftig and Birman, 1965; Clark, 1967). The small size of glaciers in the South Fork American River drainage (Wahrhaftig and Birman, 1965) on the western slope, poor preservation of glaciated features in the less competent rock types near and north of Lake Tahoe, and the documentation of the West Walker glacier as the largest glacier on the eastern side of the range (Clark, 1967) certainly suggested a decrease in glacier size north of the Yosemite region.

My reconnaissance field investigations, supplemented with work by M. Clark and A. Gillespie, indicate that the headwaters of the Mokelumne drainage were covered by a large late-Pleistocene ice field that overflowed a greater percentage of the adjoining reach of the Sierra Nevada crest than did any glacier or ice field to the south. This ice field also apparently had substantial flow into the American River and Upper Truckee River drainages to the north, and the Stanislaus River drainage to the south. Furthermore, reconnaissance mapping north of the Lake Tahoe Basin suggests that the Yuba and Feather drainages to the north also supported several large ice fields, although probably smaller than the Mokelumne ice field. Thus, although the Sierra Nevada was not mantled to the foothills by late-Pleistocene ice as John Muir proposed (Muir, 1880), Tioga-age ice fields along the crest north of Yosemite may have been more extensive and characteristic of glacial ages than has been appreciated previously. Climate apparently was severe enough during the last glaciation that equilibrium lines were below some of the major ridges comprising the northern Sierran crest.

Previous Studies

LeConte (1873, 1875) made some of the earliest observations of glacial features and evidence for multiple glaciations in the Carson drainage and in the Tahoe Basin, and offered the first hint that large ice fields may have once occupied the crest at the headwaters of the Carson River. However, he overestimated the scale of maximum glaciation when he argued that a huge “mer de glace” formerly filled the entire basin now occupied by Lake Tahoe (LeConte, 1873, p. 336). LeConte’s reconnaissance studies were followed by general mapping of the northern half of the range late in the 19th century by U.S. Geological Survey geologists, including regional observations of the extent of glacial deposits (Lindgren, 1896, 1897, 1900; Turner, 1897, 1898; Turner and Ransome, 1898).

After these early observations, there were few glacial studies in the northern half of the range until the latter half of the 20th century. In a notable exception, Blackwelder (1931) described Tioga and Tahoe moraines in the Lake Tahoe Basin and along the Truckee River in his landmark paper on glaciation of the eastern Sierra Nevada. In a brief subsequent review paper (Blackwelder, 1932), he included a map showing his inferred ice limits as far north as the North Yuba River in the Sierra Nevada during the Tahoe glaciation. The map depicts Tahoe glaciers as being restricted to valleys north of the South Fork Stanislaus River, with transfluent ice fields being notably absence. However, he did not discuss his evidence for the ice limits in this paper. McAllister (1936) emulated Blackwelder's work, mapping Tahoe and Tioga moraines along the south and west shores of Lake Tahoe.

In a later study, Curtis (1951) also followed Blackwelder's view and argued for limited glaciation of the crest in the headwaters of the Carson and Mokelumne drainages, with late-Pleistocene ice restricted to small valley glaciers. In the early 1960's, Birkeland (1962, 1964; Birkeland, *in* Wahrhaftig *et al.*, 1965) completed a detailed study of glacial and related deposits in the Lower Truckee River and Donner Pass area, immediately northwest of Lake Tahoe; his investigation remains the most thorough in the northern part of the range. Birkeland (1964) recognized Tioga and Tahoe moraines, and remnants of two older drifts, the Donner Lake and Hobart tills. Although he felt confident in correlating his Tioga and Tahoe tills to those mapped to the south, on the basis of topographic position, moraine morphology, boulder weathering, and soil development, Birkeland did not think correlating the older till remnants was feasible. Birkeland also described a younger set of cirque moraines, the Frog Lake Till, and a protalus rampart. He considered the Frog Lake Till to represent a stillstand or readvance during Tioga deglaciation. The fresh boulder surfaces, negligable erosion, and sparse lichens and vegetation indicated to him that the protalus rampart was Holocene and probably Neoglacial in age.

During an extensive INQUA field trip to inspect glacial deposits in the northern half of the Sierra Nevada, P. W. Birkeland, C. Wahrhaftig, R. P. Sharp, G. Curtis, among others (Wahrhaftig *et al.*, 1965), described evidence of glaciation at several sites that are not reported elsewhere, especially in and adjacent to the Lake Tahoe region. In general, they relate strong evidence for a two-fold Tioga-Tahoe sequence, with scattered remnants of earlier glaciations.

More recently, Mathieson (1981) and Durrell (1987) conducted fairly extensive studies of the glacial record in the Sierra Buttes region, Middle Fork Feather River. Their mapping demonstrated a complex history of glacier advance and retreat, probably dating back to at least the penultimate glaciation (marine isotope stage 6). Mathieson (1981) did not correlate the deposits to the classic four-fold Sierran sequence; instead he divided the them into pre- and post-Sangamon events on the basis of relative weathering estimates, with two separate glaciations during each event. As with Blackwelder's (1932, Plate II) earlier mapping, Mathieson's and Durrell's mapping implies that glaciers in this area were confined to the valleys rather than being outlet glaciers of a larger ice field (Mathieson, 1981, Fig. 3b,c; Durrell, 1987, Fig. 158). Without describing the evidence, Durrell mentions several locations occupied by small glaciers further north along the Sierran crest, reaching as far north as Butt Creek at the northern end of the range (Durrell, 1987, p. 277). This brief description and the regional survey and reconnaissance study by Wahrhaftig and Birman (1965) are the only published acknowledgments that glaciers may have existed in the region between the Sierra Buttes and Lassen Peak.

In the Lassen Peak area, Crandell (1972) initially mapped four ages of deposits he correlated with Tioga, early Tioga, Tahoe, and pre-Tahoe stades. Gerstel (1989) later refined the age of these deposits on the eastern side of the mountain, restricting the Tioga-equivalent advance to between 31-18 ky BP based on radiocarbon dating of a 31 ky BP peat that underlies the till, and the Tahoe equivalent to early- or mid-Wisconsin time, based on weathering rinds, soil development, and moraine-dissection measurements.

Despite these previous studies, the extent of Pleistocene glaciation in the Sierra Nevada north of Yosemite remains relatively undocumented. To a large degree, this probably reflects difficult mapping conditions: lower elevations, greater precipitation, and more-easily weathered lithologies have allowed greater post-glacial erosion and more extensive forest cover than in the south, which obscure the evidence of glaciation. Probably of equal importance, the less-than-spectacular scale of the peaks, and therefore glaciation, and the predominance of thick Tertiary pyroclastic bedrock reduce the aesthetic appeal of the northern end of the range. As a result, this part of the Sierra Nevada has not fired the imaginations of geomorphologists to nearly the same degree as the more spectacular southern half of the range.

The main focus of this chapter, then, is to lay a basic framework for future detailed work on glaciation in the northern Sierra Nevada by describing reconnaissance observations and mapping of late-Wisconsin (Tioga) glaciers and ice fields that existed in the Sierra Nevada north of the Stanislaus River (Fig. 1.1, 1.2). My study is not exhaustive and not only are there sites where Pleistocene ice existed, but also sites not included here that I have missed.

METHODS

The methods emphasize documenting the spatial distribution of late-Pleistocene glaciation in the northern part of the Sierra Nevada. Mapping is based on interpretation of small-scale (1:60,000) and large-scale (1:16,000) black-and-white aerial photographs in conjunction with detailed field mapping of selected areas. I established late-Wisconsin

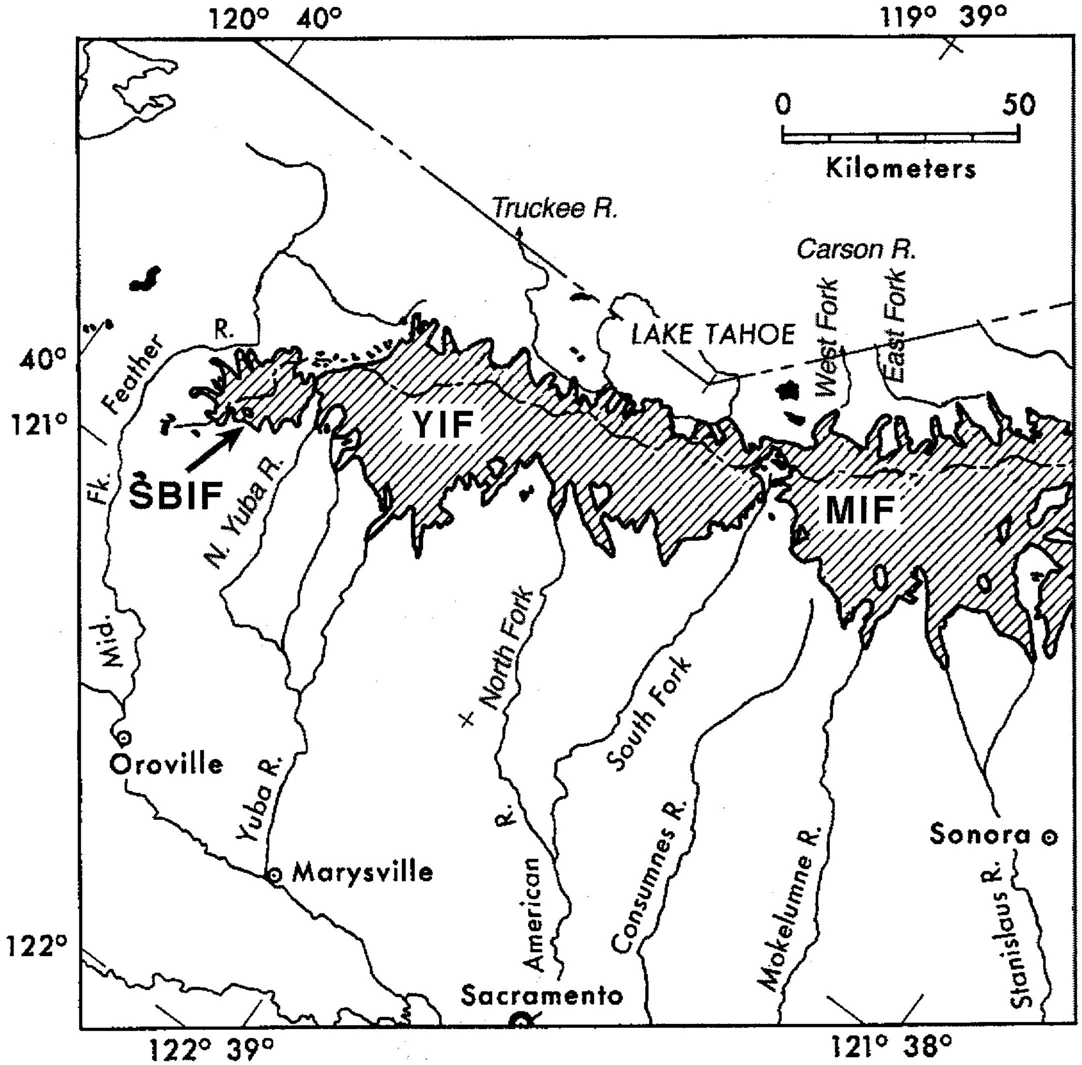


Figure 1.1: Location diagram of northern Sierra Nevada, showing approximate extent of Tioga ice fields (shaded), isolated valley glaciers (black), major drainages, and crest of range (dashed line within glaciated region). Modified from Wahrhaftig and Birman (1965). MIF, Mokelumne ice field; YIF, Yuba ice field; SBIF, Sierra Buttes ice field.

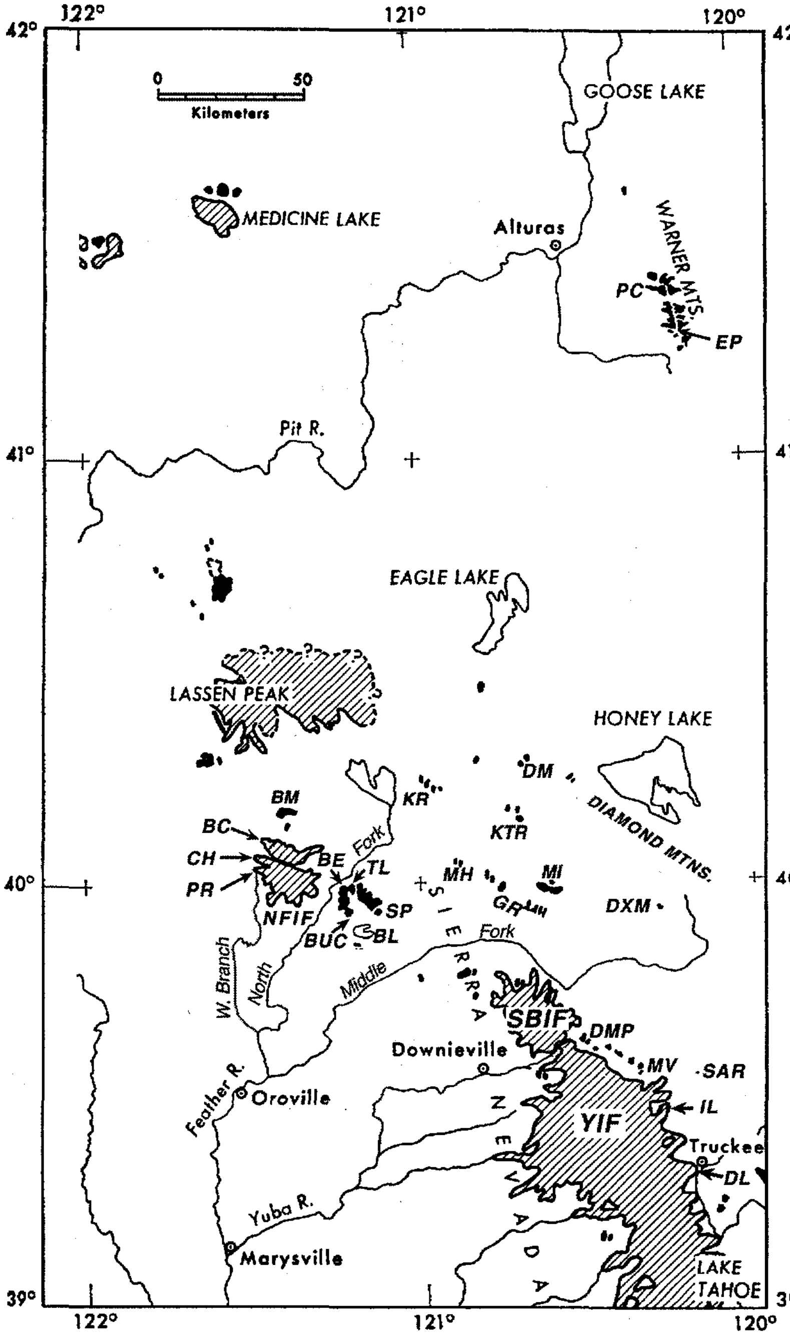


Figure 1.2: Approximate Tioga glacier limits in northeastern California (modified from Wahrhaftig and Birman, 1965). Limits of ice fields shown as shaded areas, limits of larger isolated valley and cirque glaciers shown as black areas. BC, Butte Creek; BE, Bald Eagle Peak; BL, Buck Lake; BM, Butt Mountain; BUC, Buck Mountain; CH, Coon Hollow; DL, Donner Lake; DM, Diamond Mountain; DMP, Deadman Peak; DXM, Dixie Mountain; EP, Eagle Peak; GR, Grizzly Ridge; IL, Independence Lake; KR, Keddie Ridge; KTR, Kettle Rock; MH, Mount Hough; MI, Mt. Ingalls; Maiden Valley; NFIF, North Feather ice field; PC, Pine Creek; PR, Philbrook Reservoir; SAR, Sardine Peak; SBIF, Sierra Buttes ice field; SP, Spanish Peak; TL, Three Lakes; YIF, Yuba ice field.

ice limits within the accumulation zones of glaciated regions based on trimlines and the morphology of overridden or ice-abraded ridges versus nunataks, and on distribution of erratics, mainly granitic boulders on volcanic bedrock.

Where possible, I assigned lower ice limits within the ablation zone based on location and continuity of late-Wisconsin lateral and terminal moraines. I have focused on mapping the upper limits of morainal deposits, especially those locations where lateral moraines of adjacent drainages diverge at the margins of an ice field, because these help define the lower limit of moraine deposition and confluent ice. This limit theoretically coincides with the equilibrium line of an ice field or glacier. A main advantage of using lateral moraines to define the former ELA is that their upper limit should be independent of the hypsometry of a glacier, which can substantially effect other methods (e.g., those based on accumulation-area ratios, terminus-to-headwall-altitude ratios). On the other hand, using lateral moraines to define ELAs has a couple of potential problems (e.g., Meierding, 1982): (1) poor preservation of the upper limit of lateral moraines because of erosion or discontinuous deposition near the paleo-ELA; and (2) time-transgressive lateral moraine formation as the glacier retreats. The first problem will lead to underestimates of the ELA, whereas the latter problem will result in lateral moraines that reach above the ELA of the maximum advance, and therefore give an overestimate of the ELA. Although these problems potentially decrease the accuracy of the method, they are not intractable. Divergent outlet glaciers of an ice field are particularly useful because they clearly mark the true upper limit of deposition in an ice field, especially when preserved on flat surfaces where post-deposition erosion can be discounted. Distinct lateral moraine crests that are not joined by recessional moraines or that form a single distinct crest continuous with the maximum terminal moraine suggest that the moraines formed during the maximum glacier advance and were not extended upslope during retreat of the glacier. I have applied these criteria in my study.

Lateral moraines are particularly important for estimating the ELAs of ice fields, because some methods of estimating ELAs (e.g., using the accumulation-area ratio method) are not possible without detailed knowledge of flow lines and ice divides, both of which require detailed mapping beyond the scope of my study. Other methods (e.g., using cirque-floor altitudes, terminus-to-headwall altitude ratio) are inappropriate to the physiography or hypsometry of ice sheets. I therefore rely on the highest occurrence of lateral moraines as a proxy indicator of the ELA in my study of the northern half of the range, assuming regional synchrony.

I distinguished remnants of the Tioga stade from those of older advances based on traditional relative-dating techniques (e.g., Blackwelder, 1931; Sharp, 1960; Birkeland, 1964; Clark, 1967): 1) spatial and relative positions of deposits; 2) preservation of original morphologies; 3) soil development; and 4) local qualitative assessments of boulder weathering and boulder frequency. Throughout much of the northern half of the range, Tioga moraines are generally quite distinct from older deposits, having substantially less soil development, less dense forest cover and denser shrub cover, "young" morphologies with slopes that are commonly only slightly lower than the angle of repose, with internal drainage, minor postglacial incision, and abundant surface boulders (e.g., Wahrhaftig and Birman, 1965, p. 307).

Semiquantitative measurements on moraine boulders (e.g., weathering, frequency, and relief) may not be useful as a regional correlation tool throughout much of the northern end of the range because the bedrock in many areas was previously, and variably, weathered during the Tertiary. In particular, deeply weathered granitic surfaces in some areas, which had been buried by Tertiary pyroclastic deposits, have since been reexposed by Pleistocene glaciers. This heterogeneous prior weathering violates one of the basic assumptions needed to correlate moraines between different regions using boulder weathering, namely that glacially plucked rock weathers from chemically unaltered initial states. Such pre-glacial weathering is particularly evident in the Mokelumne and Carson drainages, where large expanses of glaciated granodiorite bedrock in the former glacier accumulation zones remain deeply weathered. The weathered zones, sometimes reaching structural thicknesses as great as 50-100 m, occur below a regional Tertiary erosion surface that was buried by thick Miocene volcaniclastic deposits (Fig. 1.3) and is now being exhumed (Fig. 1.4; Curtis, 1951). Granitic bedrock near this exhumed contact, which was demonstrably overrun by large Tioga glaciers (e.g., at Ebbetts Pass), is commonly quite grusy, with no preservation of striae or glacial grooves. Erratics within the Tioga ice limits are often similarly weathered, with xenolith relief of 10 cm or more (Fig. 1.5). It is unlikely that such rapidly crumbling rocks were pristine when initially deposited. Furthermore, in the northern end of the range, much of the bedrock is volcaniclastic or sedimentary, and regionally heterogeneous, resulting in many different weathering rates of plucked clasts. Such extreme variability of the weathering rates generally limits the usefulness of boulder weathering assessments to establishing local relative sequences rather than regional correlations.

Because the trimlines and moraines of the maximum Tioga advances are generally distinct morphologically, both on air photos and on the ground, I used the Tioga advance as the baseline by which to compare earlier or later advances. However, the low elevation of most of the peaks, especially north of Lake Tahoe, apparently precluded post-Tioga (Recess Peak, Matthes) glacier advances, although local protalus ramparts may have formed penecontemporaneous with these advances (Birkeland, 1964). Pre-Tioga advances are also only locally evident. To a large degree, this reflects the apparently poor preservation of moraines composed of reworked volcaniclastic bedrock, as well as the poor exposure due to forest cover. Older deposits may exist in many of the areas where I have not mapped them, but it will take more-detailed mapping to distinguish them.

Figure 1.3: Contact between weathered granitic bedrock and overlying late-Tertiary volcanic Mehrten Formation. Contact is typically low-relief, with abundant petrified wood indicating buried forests; granitic rock below contact is commonly extensively weathered, up to depths of 50-100 m. Trees in center along contact are ~2 m tall. View to SW.

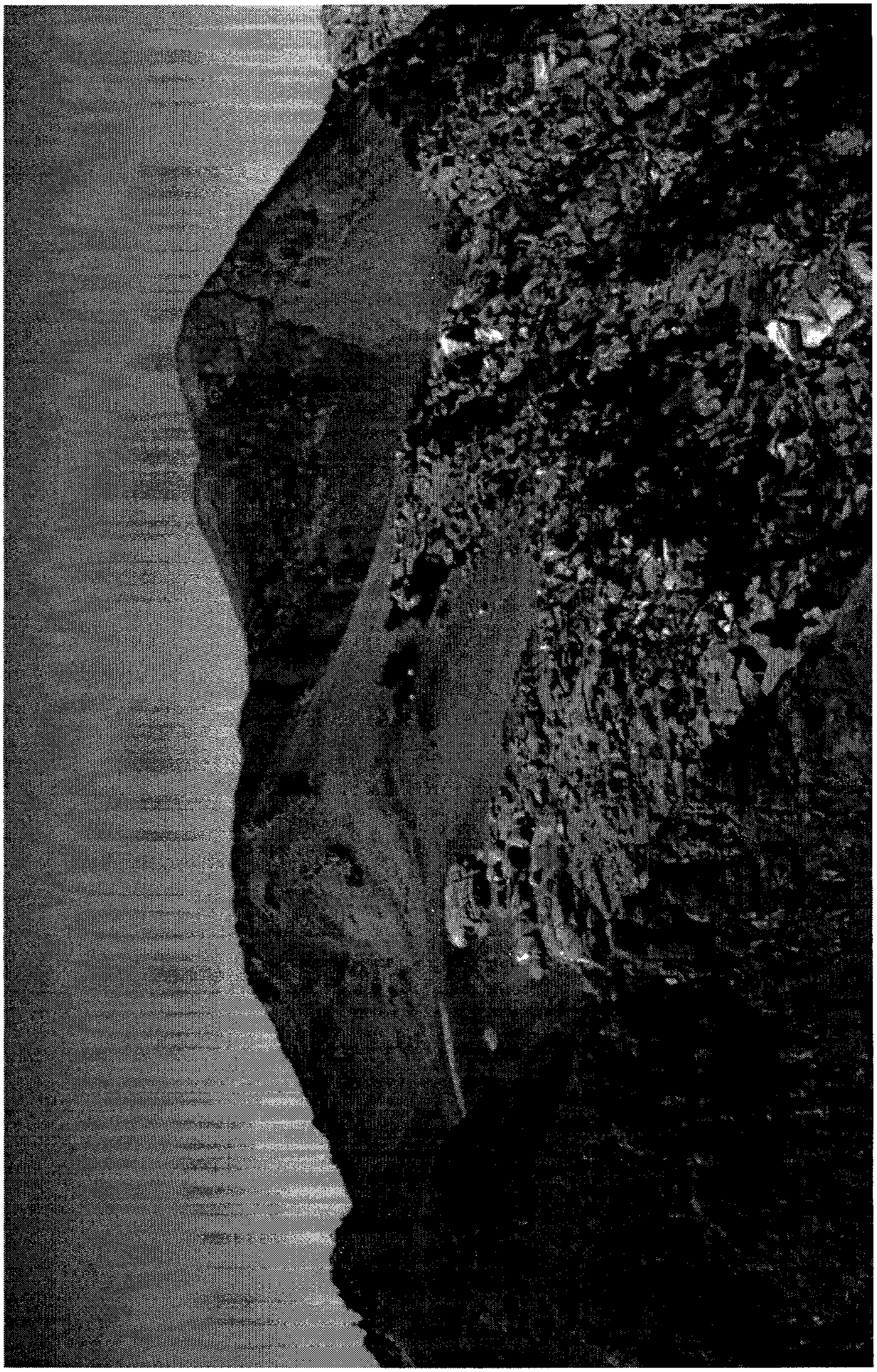


Figure 1.4: Oblique aerial view to north of North Fork Mokelumne River, showing broad, low-relief granitic plateau being exhumed, primarily by Pleistocene glacial erosion. White lines show divides with adjoining drainages; most divides formed by ridges of late-Tertiary volcaniclastic deposits that still locally bury the Tertiary surface. Lake Alpine is ~1.2 km long.

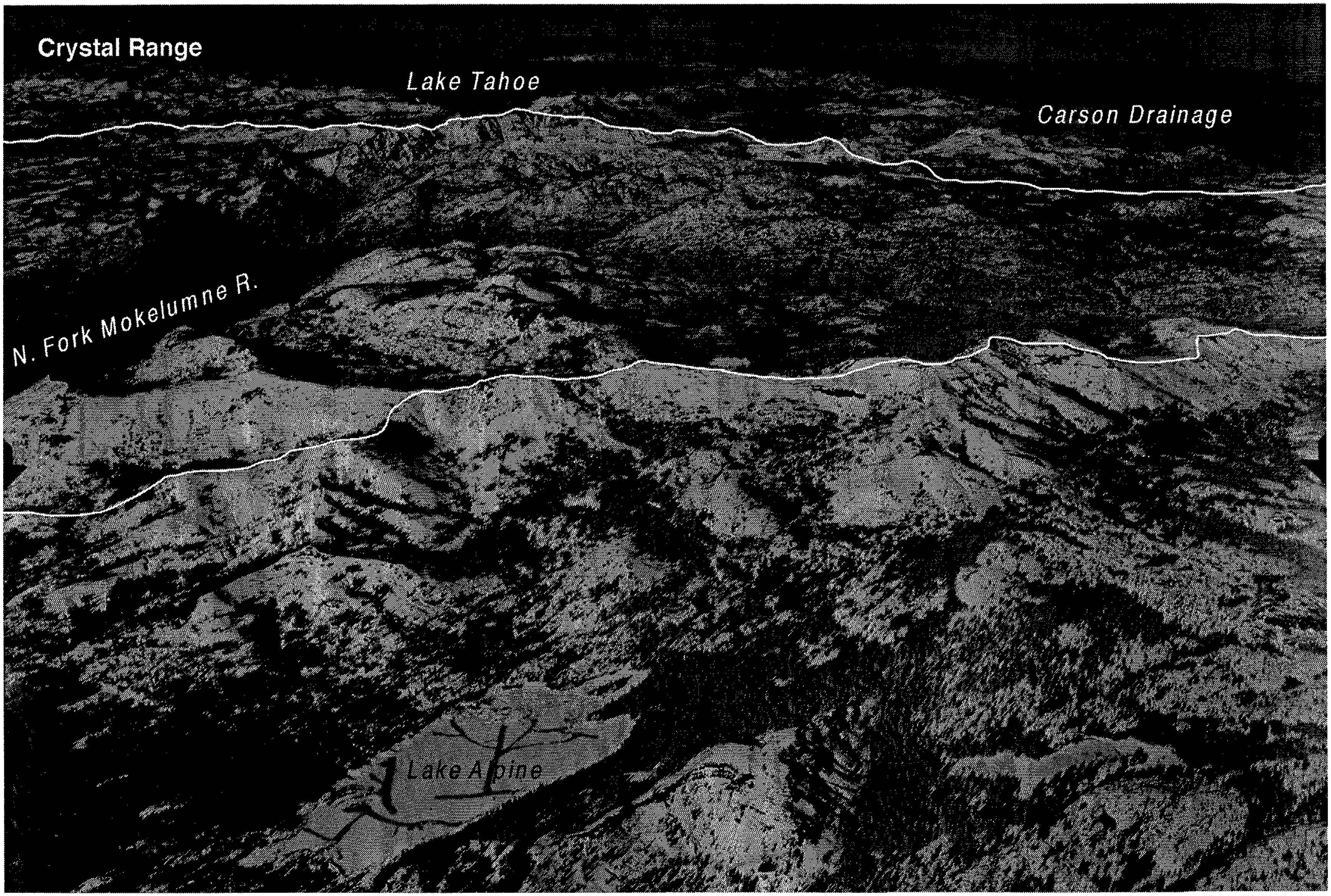
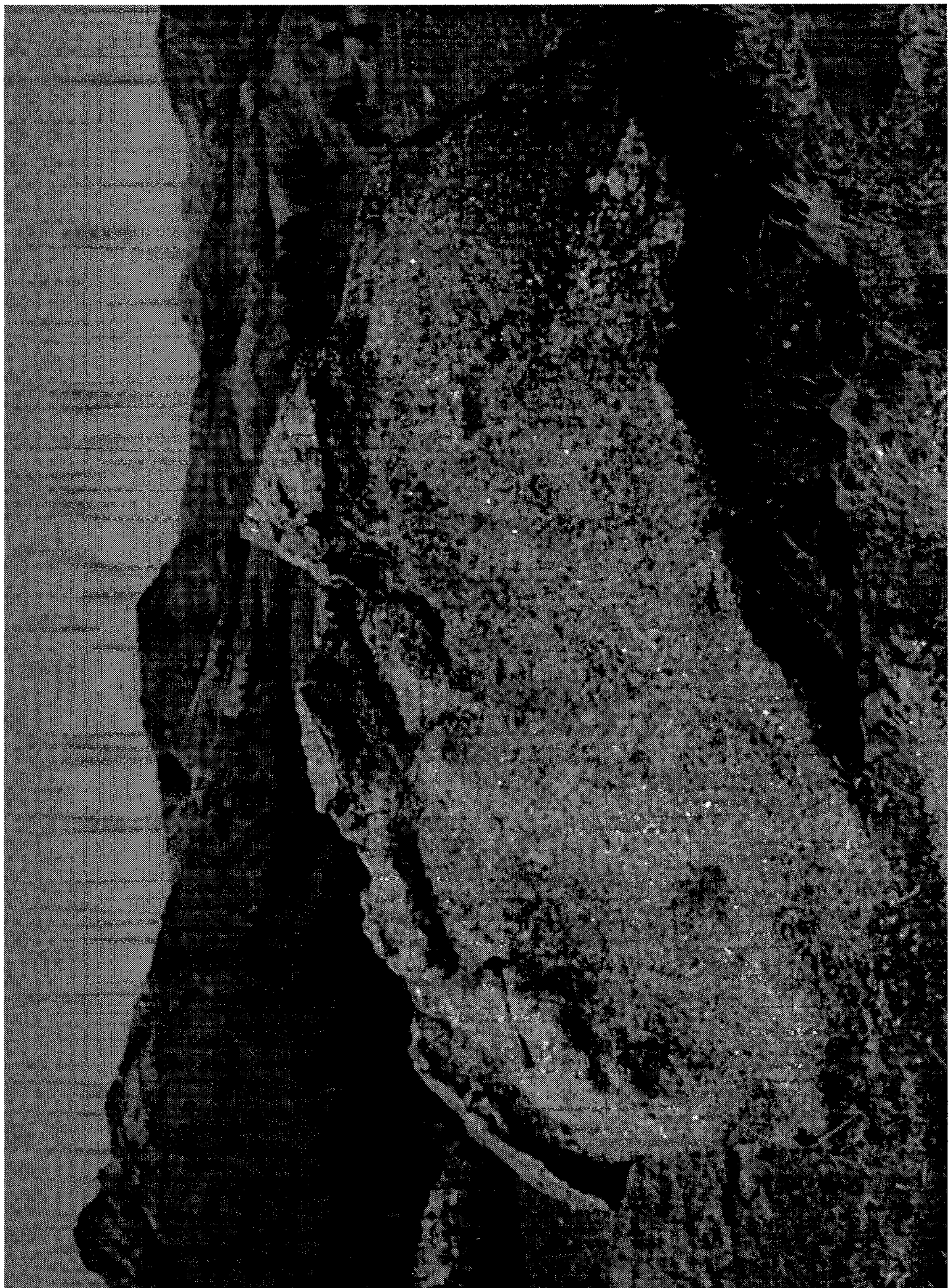

Crystal Range
Lake Tahoe
Carson Drainage
N. Fork Mokelumne R.
Lake Alpine

Figure 1.5: Tioga-age granitic erratic in Mokelumne drainage near Sierra Nevada crest, showing extensive post-depositional erosion (>10 cm) as a result of pre-glacial weathering of bedrock from which boulder was plucked. Note rock hammer for scale on left side of boulder. View is to northwest; Round Top is peak on left skyline.



Blackwelder (1931) referred to the large moraines on the southwest side of Lake Tahoe when defining the Tioga and Tahoe glaciations, with the Tioga advance representing the most recent maximum advance. Birkeland (1962, 1964) and Durrell (1987) considered that correlating the youngest glacial-maximum deposits north of Lake Tahoe to the Tioga advance was feasible and straightforward. I adopt their and Blackwelder's reasoning, and tentatively correlate deposits of the most recent glacial maximum to the Tioga stade. Correlations of older deposits (with Tahoe or pre-Tahoe advances) are probably not justified, considering the confusion about timing of pre-Tioga advances in the range (e.g., Phillips *et al.*, 1990; Gillespie, 1991), and the reconnaissance nature of my study.

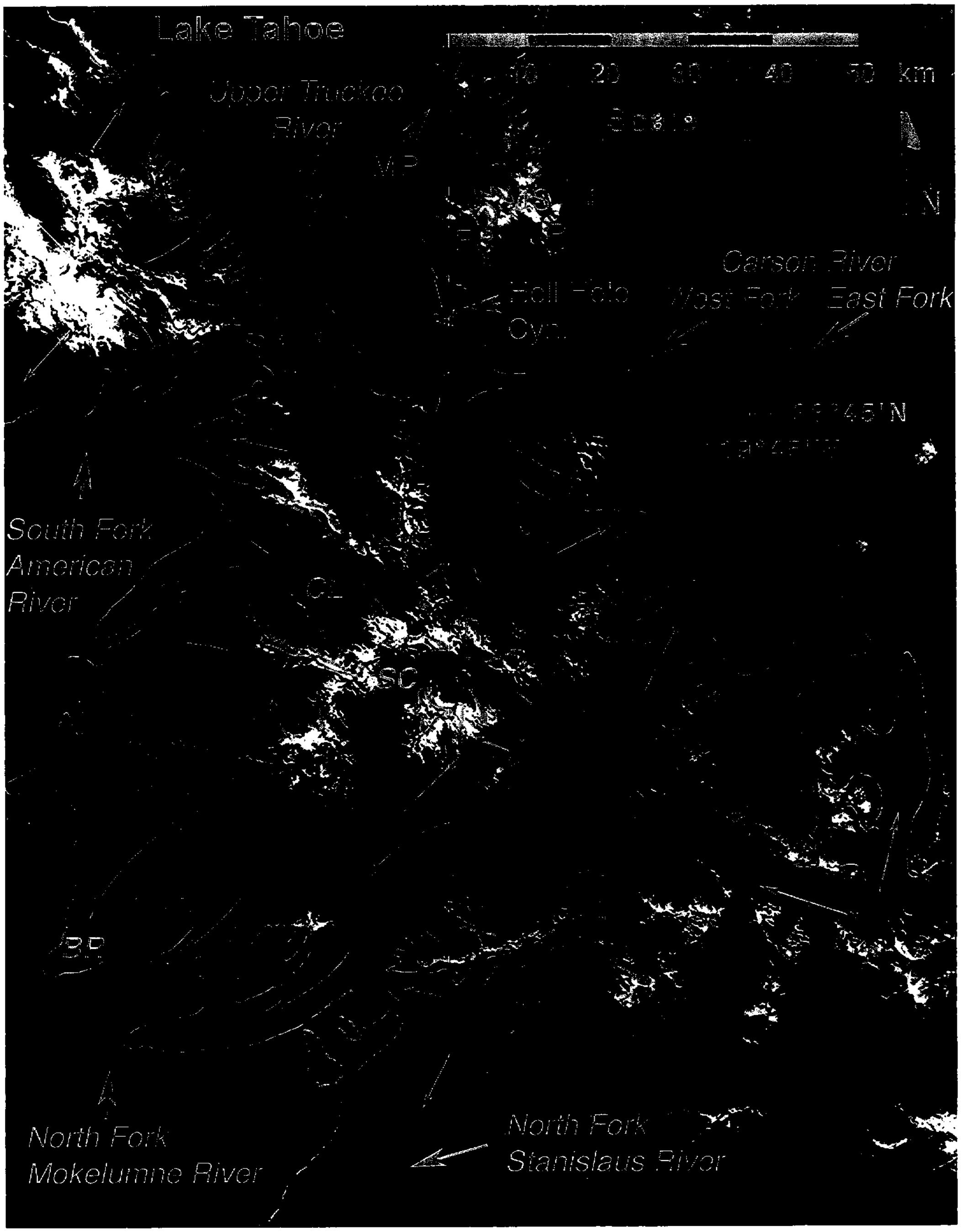
Flow directions in an ice field are primarily a function of the topography of the ice surface and of the emergent bedrock topography. In the Mokelumne ice field, I have established flow lines, based on striae, canyon shapes, trimlines, and stoss-and-lee topography, as an independent measure of the extent of glacier ice. I did not map most other ice fields in enough detail to establish exact flow lines.

RESULTS

Mokelumne Ice Field

The Tioga-age Mokelumne ice field (MIF) was substantially larger than the isolated valley glaciers indicated by Blackwelder (Blackwelder, 1932, Plate II) and Curtis (Curtis, 1951) in the region, and instead was closer in scale to that mapped for the Tahoe advance by Wahrhaftig and Birman (Fig. 1.1; Wahrhaftig and Birman, 1965). It covered much of the Sierran crest along the headwaters of the Mokelumne, and overflowed several major drainage divides (Fig. 1.6). Ice from the MIF flowed south into the North Fork Stanislaus River, north into the Silver Fork of the American River, and east into the East and West forks of the Carson River. An icefield in the headwaters of Caples Creek

Figure 1.6: Landsat TM image of Sierran crest south of Lake Tahoe, showing Tioga ice limits of Mokelumne ice field. Short arrows show flow directions from striae and stoss and lee forms; long thin arrows show regional flow directions from topography and trim lines. "D" shows locations of ice domes. BR, Rear River Reservoir; CL, Caples Lake/Caples Creek; FP, Freel Peak; JP, Jobs Peak; MP, Monument Peak; JS, Jobs Sister; SC, Summit City canyon.


Lake Tahoe
Upper Truckee River
10 20 30 40 50 km
Scale
N
MP
Hell Hole Cyn.
Carson River
West Fork
East Fork
South Fork American River
CL
SC
BP
North Fork Mokelumne River
North Fork Stanislaus River

(Silver Fork American River) that was continuous with the MIF also flowed northward into the Upper Truckee River, as well as eastward across Carson Pass into the West Fork Carson River. Trim lines, striae, erratics, and canyon morphology indicate that several ice domes were located within the MIF, reaching altitudes of ~2900 m. These domes resulted in ice flow over the bedrock drainage divides and out of the Mokelumne drainage.

Locally, however, some ice from the adjacent drainages probably flowed into the Mokelumne drainage. A small amount of ice from upper Caples Creek drainage probably flowed into the much deeper Summit City Canyon in the Mokelumne drainage (Fig. 1.6). Similarly, a small tongue of ice apparently flowed from the North Fork Stanislaus drainage near Lake Alpine and joined or nearly joined the main MIF outlet glacier in the canyon of the North Fork Mokelumne River (Fig. 1.6).

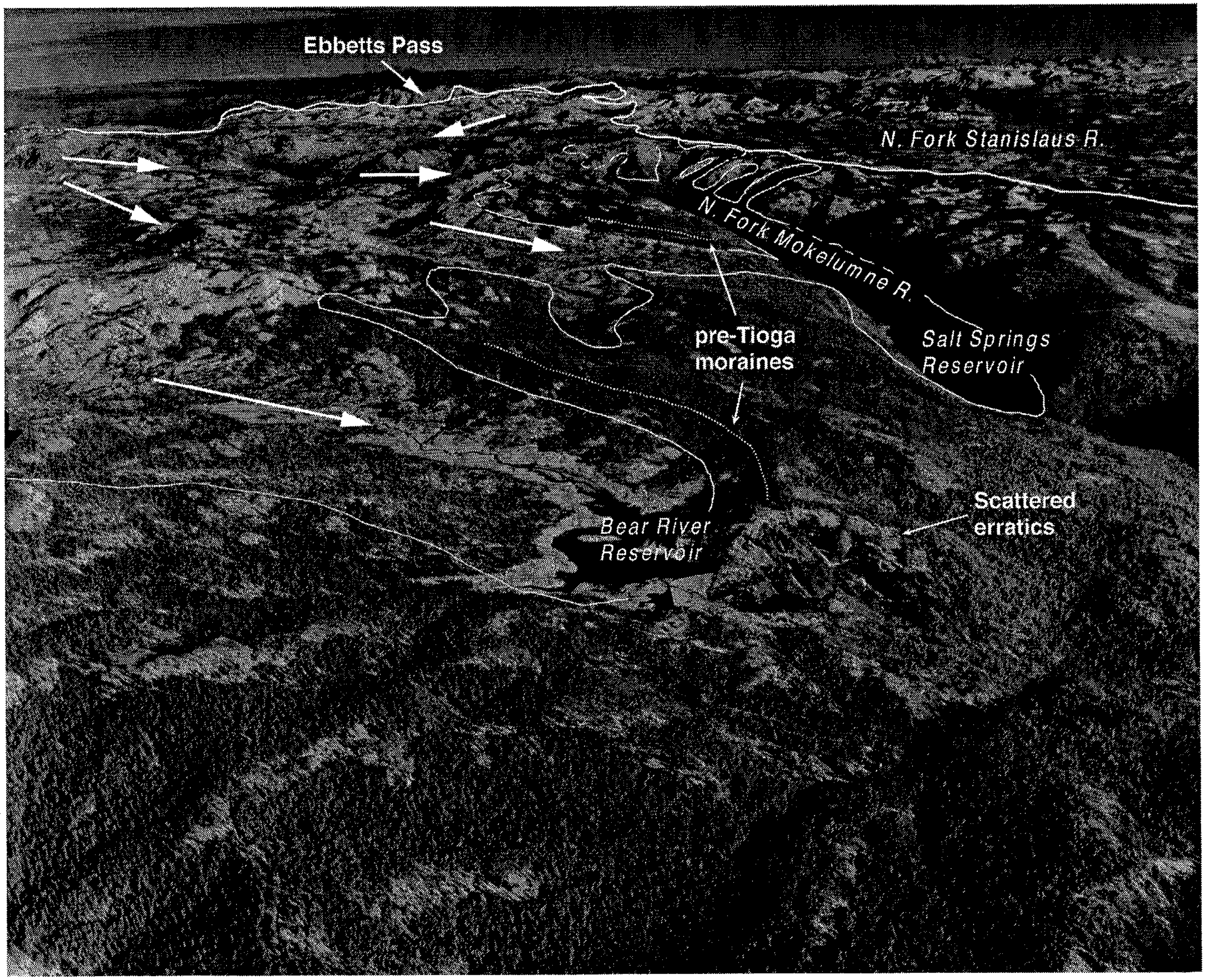
Within the limits of the MIF, 16.5 km of the Sierran crest was ice-covered, which is equivalent to 51% of the crest measured through the area. The length of the divide between the North Fork Mokelumne and South Fork American rivers that was glacierized was 22.5 km, roughly 80% of the divide within the ice field, whereas about 16 km, or 44%, of the North Fork Mokelumne/North Fork Stanislaus divide was glacierized. Although the absolute lengths of the crest and divides covered by transfluent MIF ice are not as great as those attained by the Tuolumne ice field, the percentages of areas covered are equivalent, and greater than other ice fields in the Sierra Nevada (M. Clark, personal communication, 1991). Thus, despite relatively low crestal altitudes (most below 2750 m (9000 ft)), the MIF was one of the most continuous ice fields in the range.

Tioga moraines in the Mokelumne drainage have generally lower-volume and are less well-preserved relative to moraines in the eastern Sierra Nevada and the Tahoe basin, although they are well-formed where they do occur. This situation may in part reflect the scant sources of supraglacial debris from subaerial rockfall in the ice field area. Most debris in the ice must have been plucked from the base of the glacier, therefore reducing the debris content relative to glaciers with extensive rockfall sources. Despite their irregular occurrence, the highest Tioga-maximum lateral moraines in the Mokelumne drainage have remarkably consistent altitudes of about 2100-2300 m, indicating that this altitude was the approximate Tioga ELA for the drainage. This altitude places much of the large plateau in the headwaters of the North Fork Mokelumne River above the ELA (Fig. 1.7), and helps explain the extent of ice in this area of moderate relief. Highest occurrences of lateral moraines in the West Fork Carson drainage east of the crest are near 2450 m, suggesting an ELA gradient rising about 17 m/km to the east-northeast. This gradient is slightly higher than that estimated by Burbank (1991; 13 m/km) for Tioga glaciers in the southern Sierra Nevada, but very close to his estimate based on modern glaciers (16 m/km). My estimate is essentially indistinguishable from Burbank's within the error limits of the methods (~100-150 m; Meierding, 1982).

Pre-Tioga moraines from earlier MIFs occur outside Tioga ice limits in the West and East forks of the Carson River, but are less evident west of the crest. Pre-Tioga lateral moraines are present 50-100 m above Tioga maximum moraines south of Lower Bear River Reservoir and south-southwest of Mokelumne Peak. Scattered, highly weathered erratic boulders are perched locally on bedrock pedestals outside the pre-Tioga moraines (e.g., on the bedrock knob forming the southern buttress of the Lower Bear River Reservoir dam; Fig. 1.7), providing evidence for a still older, more extensive glaciation.

The western extent of the MIF along the drainage divides between the Mokelumne River and the American and Stanislaus rivers is unclear. Dense forest cover and less-competent, rapidly erodable bedrock (Tertiary pyroclastic rocks rather than granite), hide evidence of glaciation and combine to obscure the former ice limits in these areas.

Figure 1.7: Oblique aerial view to east of North Fork Mokelumne River, showing approximate limits of Tioga ice (thin lines; arrows show approximate flow directions) feeding off broad plateau. Thin dotted lines show location of two prominent pre-Tioga (Tahoe) moraines above Tioga moraines. Bold white line shows divides with adjoining drainages, much of which was covered by the ice field. Note low relief of most of plateau (compare to Figure 4). Water in Bear River Reservoir is ~1 km wide just above dam.


Ebbetts Pass
N. Fork Stanislaus R.
N. Fork Mokelumne R.
pre-Tioga
moraines
Salt Springs
Reservoir
Bear River
Reservoir
Scattered
erratics

Truckee and American River Drainages

Evidence for glaciation is abundant in the mountains surrounding Lake Tahoe, particularly along the main crest west of the lake. A large Tioga ice field existed in the mountains between the South and North forks of the American River; large outlet glaciers flowed onto the plateau between the North and South forks to the west, and into Lake Tahoe to the east. The glaciers flowed from altitudes as high as ~2920 m and terminated at ~1850 m. The moraines in the Tahoe basin are particularly well developed; those of glaciers that flowed off the west side of the crest are similarly abundant, though less apparent in the forest cover. These deposits were previously mapped in reconnaissance (Lindgren, 1896; Lindgren, 1897; McAllister, 1936; C. Wahrhaftig *in* Wahrhaftig *et al.*, 1965; Loomis, 1983); only those in Desolation Wilderness at the southern end of the ice field have been mapped in detail (Whiting, 1986). Tioga lateral moraines on both sides of the crest in this area occur up to ~2300 m.

In contrast to glaciers on the main crest, those in the Carson Range east of the Lake Tahoe were generally restricted to valleys on north-facing slopes. Small Tioga valley glaciers (~4-5 km in length) flowed off the north sides of Freel Peak, Jobs Sister, and Freel Meadow at the south end of the Carson Range. The Hell Hole glacier, which flowed out of the cirque north of Freel Meadow, formed a particularly well developed sequence of Tioga recessional moraines that form one inside the other nearly from the terminal moraine to the cirque headwall. Possible moraines of pre-Tioga advances for each of these glaciers extend downstream from the Tioga ice limits. Smaller cirque glaciers may have existed on the north north-facing cirques of Monument Peak and Jobs Peak, but field mapping is needed to verify this. Tioga lateral moraines in this area extend up to altitudes of about 2550 m.

The northern end of the Carson Range contains both Tioga and pre-Tioga deposits (Thompson and White, 1964; Wahrhaftig *et al.*, 1965) from apparently small ice fields near Mt. Rose. However, no one has yet studied this glacial sequence in detail.

In contrast, Birkeland (1962, 1964; Birkeland, *in* Wahrhaftig *et al.*, 1965) completed probably one of the most detailed studies of glacial deposits in the Sierra Nevada along the Lower Truckee River, from Lake Tahoe to Reno, including major tributaries as far north as Prosser Creek. Birkeland mapped Tioga till and outwash terraces, as well as till of the Tahoe and two older glaciations. He distinguished the deposits on the basis of soil formation and other relative weathering criteria. Although Birkeland (1962, p. 91) described wood and charcoal exposed in a Tioga end moraine of the Donner Lake glacier near Truckee High School, he did not sample or date it. The exposure has since been destroyed.

Yuba Ice Field

Although Birkeland's work hints at it, my reconnaissance studies and recent unpublished work by J. Yount and D. Harwood (personal communication, 1992) indicate that a large ice field covered much of the Sierran crest between the headwaters of the North Fork American, North Fork Yuba, Truckee, and Little Truckee rivers. The Yuba ice field (YIF) produced many large outlet glaciers flowing west, east, and north (Fig. 1.1, 1.2). Both Donner and Independence Lakes are dammed by Tioga end moraines from this ice field. Although not as big or continuous as the MIF, there was substantial ice flow across the crest in the YIF. Because many of the nunataks in this ice field consist of distinct volcanic and shallow intrusive rock types (in contrast to the granitic bedrock that dominates the valley floors), a detailed study of provenance of transported boulders and striae attitudes should allow one to determine the flow patterns of the former ice field accurately.

The northern limit of the YIF appears to have been a tongue of ice that flowed down the Middle Yuba River past Jackson Meadow Reservoir, terminating at ~1675 m where the river turns to the west from a northwest heading. A significant tongue of this glacier apparently split and flowed over a low pass into Milton Creek and the North Fork Yuba River drainage. Absence or poor preservation of moraines make the extent of this ice tongue unclear, but valley morphology suggests that it may have reached the main North Fork Yuba River. Eastern limits of the YIF are indicated by large moraines immediately downstream from the lower ends of Independence and Donner Lakes (Birkeland, 1964). Moraines delimiting the western limits of the ice field, however, are indistinct and have received little attention. The southern end of the ice field merged with ice flowing out of the mountains on the west side of Lake Tahoe. The YIF formed the northernmost extension of a continuous Tioga ice field that stretched northward from the southern Sierra Nevada (e.g., Wahrhaftig and Birman, 1965, Fig. 2), with the possible exception of a small gap in the headwaters of the South Fork American River. However, the distinct gap with the Sierra Buttes ice field at the northern end of the YIF was small (<2-3 km), and may have been absent during earlier, more extensive glaciations.

Ice in the YIF flowed from altitudes near 2750 m to about 1830 m. Tioga lateral moraines occur up to about 2300 m on the east side of the ice field; those on the west side of the ice field are not well expressed and have not yet been mapped.

Several ages of pre-Tioga till are evident at the margins of the YIF. Those mapped by Birkeland (1962, 1964) at the southeastern margin of the ice field are the best documented. Reconnaissance inspection of the morphology and boulder weathering of the moraines surrounding Independence Lake suggest that a sequence similar to Birkeland's exists there. J. Yount (personal communication, 1992) has recently mapped Tioga and older deposits to the west of the crest in the North Fork American River, and in the Little Truckee River.

Sierra Buttes Ice Field

Immediately north of the northern limit of the YIF, a small but well developed ice field occupied the Sierran Crest for about 20-25 km north of Sierra Buttes (Fig. 1.1, 1.2). The deposits from ice flowing off the northeastern slope of this ice field, informally named for the Sierra Buttes, have been mapped in detail by Mathieson (1981a, 1981b) and Durrell (1987). Their studies imply that the ice forming these deposits originated as a series of separate valley glaciers and small ice fields that did not cover the crest, rather than a transfluent ice field (Mathieson, 1981a, Fig. 3b,c; Durrell, 1987, Fig. 158). However, rounded stoss-and-lee topography and striated bedrock indicate that ice covered significant parts of the headwaters of this reach of the crest during the Tioga stade. Tioga glaciers on the northeast slope of the ice field flowed from altitudes near 2300 m to altitudes of ~1650 m, with lateral moraines occurring up to about 2000 m. The western and northern limits of this ice field are still undefined, although valley morphology suggests that the glaciers were smaller than those that flowed to the northeast.

Spanish Peak Ice Field

The crest of the Sierra Nevada did not support any significant late-Pleistocene ice fields for about 35-40 km northwest of the Sierra Buttes ice field. Fifteen kilometers west of Quincy, however, a small Tioga ice field occupied the northeast side of the ridge trending northwest from Spanish Peak (Fig. 1.2). The steep escarpment on which this ice field formed apparently created an efficient trap for snow blowing off the broad, gently southwest-dipping side of the ridge; the southwest slope does not show evidence of having been glaciated in the late Pleistocene.

Several glaciers flowing out of the Spanish Peak ice field, from an altitude of 2070 m to altitudes of 1670 m, deposited sets of well formed moraines of Tioga age as well as probable remnants of earlier glaciations (Fig. 1.8). Tioga recessional moraines dam several small lakes, the largest of which is Silver Lake, ~1.4 km long. The accumulation zone for the ice field was semi-continuous from Spanish Peak to Lost Lake, a distance of about 6 km. The transition with the ablation zone is marked by the divergence of ice tongues and formation of medial and diverging lateral moraines. These moraines reach altitudes of 1920 m. Isolated cirque and valley glaciers occupied most of the canyons feeding off the ridge west of Lost Lake for about 5 km. In the same direction, Three Lakes is dammed by the terminal moraine of a large cirque glacier that flowed from ~2070 m to 1840 m (Fig. 1.2). The right-lateral moraine for this glacier is distinct and reaches to 1980 m. Other small lakes are formed behind Tioga moraines of small glaciers originating in north-facing cirques on the sides of Bald Eagle Mountain and Bucks Mountain (Fig. 1.2). The extent of these glaciers is not clear, but probably less than 1-2 km.

North Feather Ice Field

West-northwest of the Spanish Peak ice field, between the main and West branches of the North Fork Feather River, a low, north-trending plateau shows distinct evidence of supporting the northernmost sizable ice field in the Sierra Nevada. Tioga-age lateral moraines clearly define the ice limits of several outlet glaciers that flowed out of the south and west sides of an ice cap that covered the center of the unnamed ridge (Fig. 1.9). Distinct terminal moraines are only preserved on the western side of the ridge, where Philbrook Creek, Coon Hollow, and Butte Creek all display extensive Tioga end and recessional moraines. Each of these outlet glaciers was about 6-7 km long at their maximum late-Wisconsin extent. The ice cap that fed them covered all but perhaps a few small nunataks. The extent of the ice to the north and east is unclear, and would require extensive field mapping to discern because the moraines and trimlines on these slopes are

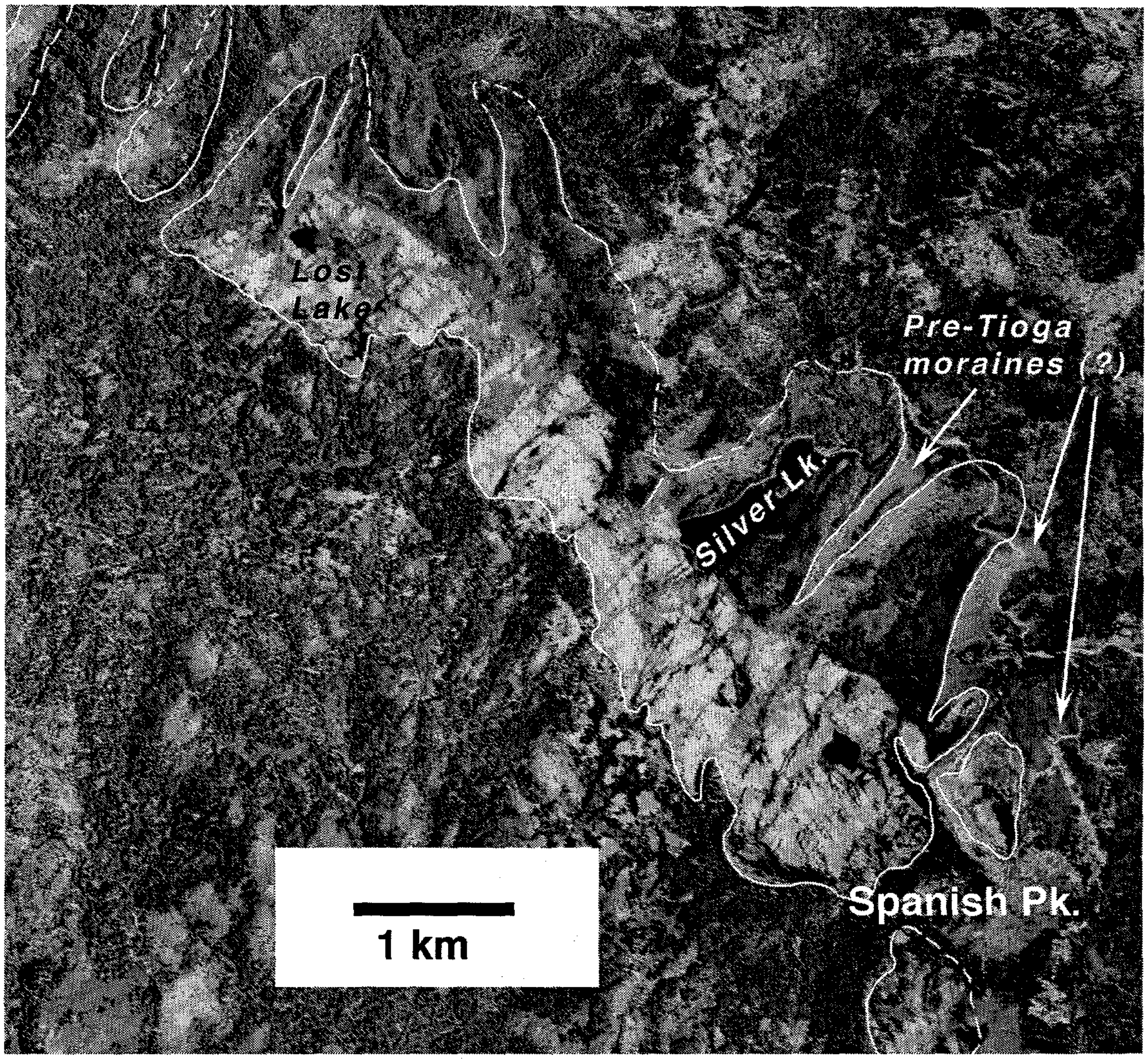


Figure 1.8: Vertical aerial photo showing approximate limits of Tioga ice in Spanish Peak ice field. North is toward top of page. Flow is to northeast across Spanish Peak escarpment. Light-colored bedrock in accumulation zone is granitic. There is no evidence of ice flowing to the southwest off the plateau.

Figure 1.9: Vertical aerial photo showing approximate southern limits of Tioga ice in North Feather ice field. North is toward top of page. Arrows show ice flow directions, based on topography and trim lines. Light-colored granitic bedrock is exposed in core of range. Note that few nunataks poked out of this ice field, making it one of the few true ice caps in the Sierra.

Table Mtn.
Big
Kimshew Cr.
1 km
Rock Cr.

apparently more subtle or poorly preserved than those to the west and south. Possibly the ice was continuous with glaciers on Butt Mountain at the north end of the ridge, but detailed field mapping is needed to test this idea.

Maximum altitudes in the center of the ice field, judged from the altitudes of possible nunataks and projection of lateral moraines, were probably 2130-2150 m, with terminal moraines at about 1650 m. Divergent lateral moraines on the western and southern slopes start at about 1830-1950 m.

Aerial photo interpretation and reconnaissance field inspection suggest that tills and moraines of at least one and probably more pre-Tioga glacier advances exist in this area. Detailed field work is needed to establish the sequence of these advances. The deposits from this ice field present an interesting location to compare relative rates of weathering of moraines composed of different lithologies, because the glacier covered several different types of bedrock, including those most common in the glaciated northern Sierra Nevada: Mesozoic granite, Tertiary volcanics, Precambrian metavolcanics, and Paleozoic marine sediments (Lyndon *et al.*, 1960; Burnett and Jennings, 1962). Such a study might enable better correlation of deposits in the central Sierra Nevada with those further to the north in the Cascade Range. However, extensive logging in the area has undoubtedly disturbed much of the surface of the moraines.

Valley and Cirque Glaciers North of Lake Tahoe

In addition to the ice fields, smaller valley glaciers and cirque glaciers occurred in scattered locations in the northern Sierra Nevada during late-Pleistocene time, typically on the north, east, or western slopes of some of the higher peaks. Most of these glaciers were less than 2-3 km long, and probably depended to a large degree on the protection afforded them by the steep slopes and deep cirques in which they typically formed. In this section I briefly describe, from south to north, the most prominent of the glaciers and their location and approximate extent. The general locations of the larger deposits are shown on Figure 1.2.

Several small Tioga valley glaciers flowed down narrow ravines on the north side of an unnamed ridge immediately south of Sierra Buttes, across the North Yuba River (39°33'N, 120°39'W). These glaciers appear to have been no more than 1-2 km long, and probably didn't reach the North Yuba River. Although the cirques are distinct, with striated bedrock, terminal moraines were not preserved on the steep slopes. The glaciers headed at altitudes of about 1950-2070 m and terminated at ~1620 m. Lateral moraines are not apparent on aerial photographs.

Sardine Peak (23 km to the east of the unnamed ridge; 120°12'W, 39°33'N), at 2479 m altitude, is high enough to have supported glaciers and may have had a small one on its eastern slope, but moraines and cirque morphologies are not evident.

Cirques and moraines indicate that glaciers 1-3 km long occupied both sides of the northwest-trending ridge between Deadman Peak and Maiden Valley, east of Sierra Buttes (39°35'N, 120°30' W). These glaciers all headed at about 2220-2320 m and flowed to altitudes of ~1920 m. Recessional moraines from these glaciers dam several lakes. Lateral moraines reach as high as ~2310 m.

North of Sierra Valley, cirque morphologies and high altitudes suggest that Dixie Mountain (2538 m; 39°55'N, 120°15'W) may have supported several small valley or cirque glaciers on its northeastern and southeastern slopes. Any such glaciers were probably no more than 1-1.5 km long, reaching to altitudes of about 2100 m. I did not discern moraines on aerial photographs, so my inference of Tioga glaciers here is tentative.

Mt. Ingalls (2552 m; 39°59'N, 120°37'W), a prominent peak on Turner Ridge perched above two well formed cirques, supported two medium-sized valley glaciers, one flowing out of the northwest-facing cirque, the other flowing out of the southwest-facing cirque (Fig. 1.10). The extent of the Tioga glaciers is not clear, but they don't appear to have been much longer than 2 km, reaching from 2380 m to ~2100 m. The two glaciers may have merged in their upper ends where the ridge separating the cirques has been partially eroded. Two small lakes are dammed by Tioga recessional moraines in the southeastern cirque, and small moraine-dammed meadows and bogs occur in both cirques. A prominent feature of both valleys, however, is that they appear to record at least one earlier, much more extensive glacier advance. Large lateral moraines that are much more weathered and incised, and have better-developed soils, than the Tioga equivalents bound the valley further downstream (Fig. 1.10). These moraines begin at ~2150 m and reach to altitudes of ~1650 and ~1830 m on the southeast and northwest slopes, respectively. Shallow cirques on the north side of Mt. Ingalls appear to have been occupied by glaciers during the earlier event(s), but it is not clear if they supported Tioga glaciers.

On Grizzly Ridge (~40°00'N, 120°47'W), the next prominent northwest-trending ridge to the west of Mt. Ingalls, a series of small Tioga glaciers flowed down the steep northeast slope from shallow cirques just below the crest (Fig. 1.11). These glaciers attained lengths of ~1-2 km; the longest was about 2.5 km and reached from 2250 m to 1460 m altitude. The glaciers all terminated on the slope, so end moraines are typically not well preserved, although lateral and medial moraines are often continuous and quite distinct on aerial photographs (Fig. 1.11). The lateral moraines typically reach well back towards the cirques, to altitudes of ~2070 m. Late-stage Tioga recessional moraines commonly form small flats, closed basins, or ponds in the cirque headwalls. Pre-Tioga deposits are not apparent on aerial photographs.

A westerly jog of the northwestern extension of Grizzly Ridge forms Indian Falls Ridge, which had several small valley glaciers that occupied its steep northeast slope, similar to those on Grizzly Ridge. The largest glacier flowed from Mt. Hough (2204 m;

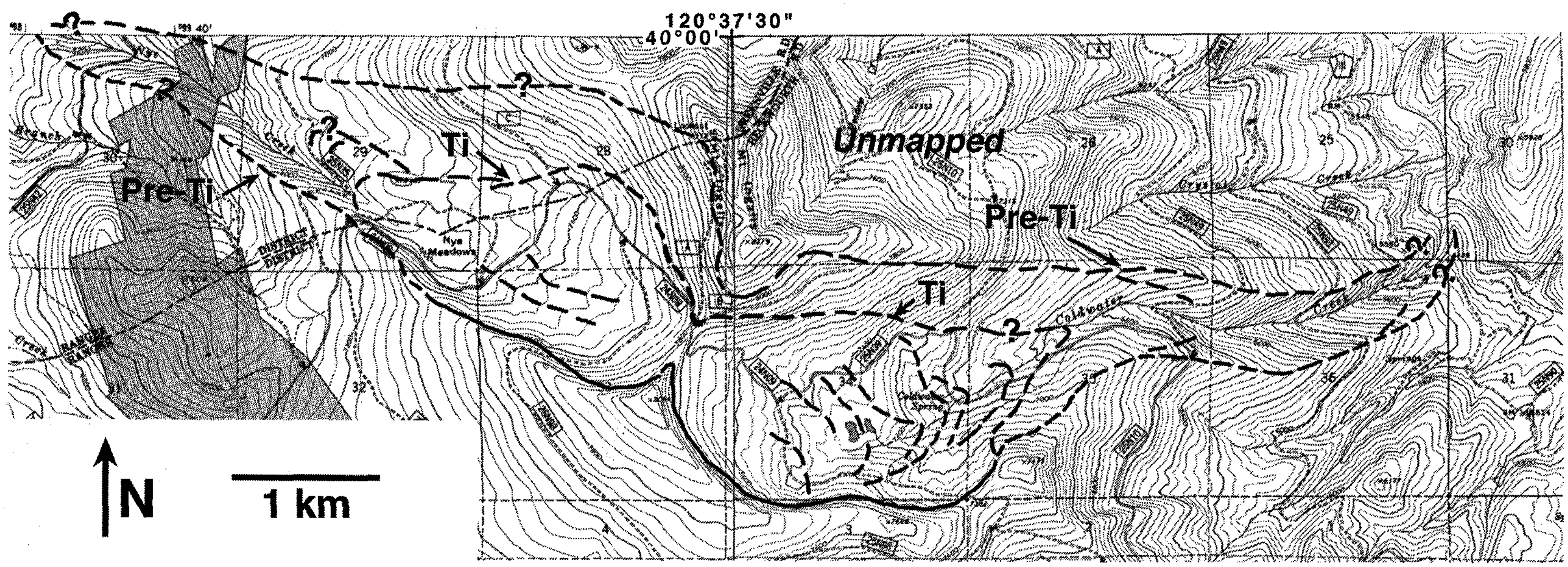


Figure 1.10: Topographic map of Tioga and pre-Tioga ice limits on eastern and western slopes of Mt. Ingalls. Poor terminal moraine preservation makes maximum extents of Tioga and pre-Tioga glaciers uncertain. Glaciers also probably occupied valleys on northern and northeastern slopes of Mt. Ingalls, but deposits there are unmapped. Turner Ridge continues to the south of Mt. Ingalls. Topography from Mt. Ingalls and Grizzly Valley, California, USGS 7.5' quadrangles. Contour interval 40'.

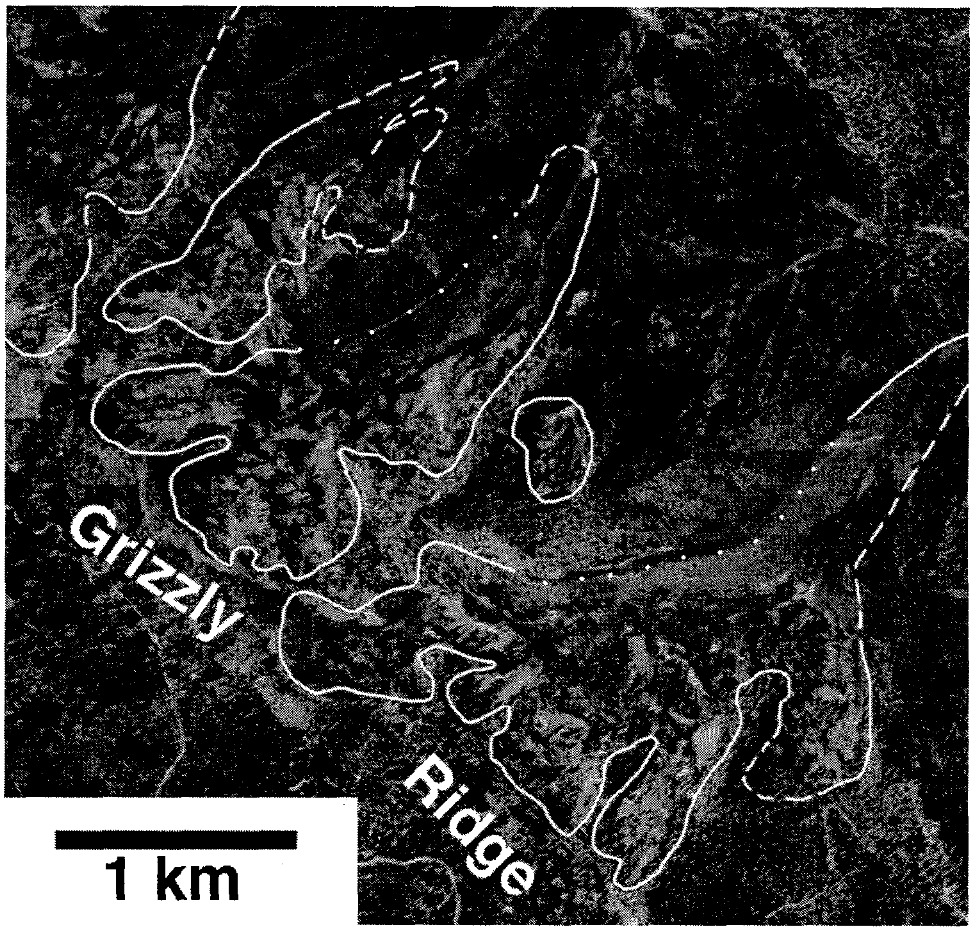


Figure 1.11: Vertical aerial photo showing limits of Tioga ice on central part of Grizzly Ridge. North is towards top of page; glaciers flowed to the northeast from top of ridge. Left-lateral moraines of two largest glaciers (dotted lines) are discrete and continuous with maximum termini, and are not joined by recessional moraines. This observation indicates that they were formed during the glacial maximum, and that the highest altitude reached by these moraines therefore approximately coincides with the Tioga-maximum ELA.

40°03'N, 120°53'W; Fig. 1.2) to altitudes of ~1340 m. As with Grizzly Ridge, terminal moraines are not well preserved; lateral moraines are also poorly expressed. Recessional moraines, however, form small benches at the heads of the cirques; one recessional moraine dams Crystal Lake beneath Mt. Hough.

South of Bucks Lake, Thompson Lake (1650 m; 39°53'N, 121°12'W) has formed behind a small moraine in a circular cirque. This Tioga glacier was quite low compared to other nearby glaciers and probably reflects enhanced accumulation in the north-facing snowdrift-fed catchment basin. Similarly, a small cirque moraine complex formed at the base of a north-facing cliff at the southwestern end of Haskins Valley (upper lateral moraines at 1700m; 39°51'N, 120°13'W) appears to be well below the regional altitude of glacier initiation as a result of enhanced wind-drift accumulation.

The northeast slope of Kettle Rock (2388 m; 40°09'N, 120°43'W), ~16 km northeast of Grizzly Ridge, was occupied by a cirque glacier (Fig. 1.2). A moraine of this glacier, which appears to be the Tioga terminal moraine, dams Taylor Lake at 2070 m. Low ridges downstream of this moraine may represent older moraines, possibly of early Tioga or pre-Tioga advances. Extensive soil and forest cover obscure their origin, however. Several small north-facing cirques ~2 km north of Kettle Rock, below Eisenheimer Peak and possibly Rattlesnake Peak, also supported small Tioga glaciers at about the same altitude.

Keddie Ridge Moraines

Keddie Ridge and Keddie Peak (2286 m; 40°13'N, 120°58'W), a northwest-trending ridge about 25 km northwest of Kettle Rock, supported several small Tioga cirque/valley glaciers on its northeastern slope (Fig. 1.2, 1.12). The maximum terminal moraines are indistinct, with the most prominent moraines being what appear to be recessional moraines that dam several lakes near the cirque headwalls. The largest of the

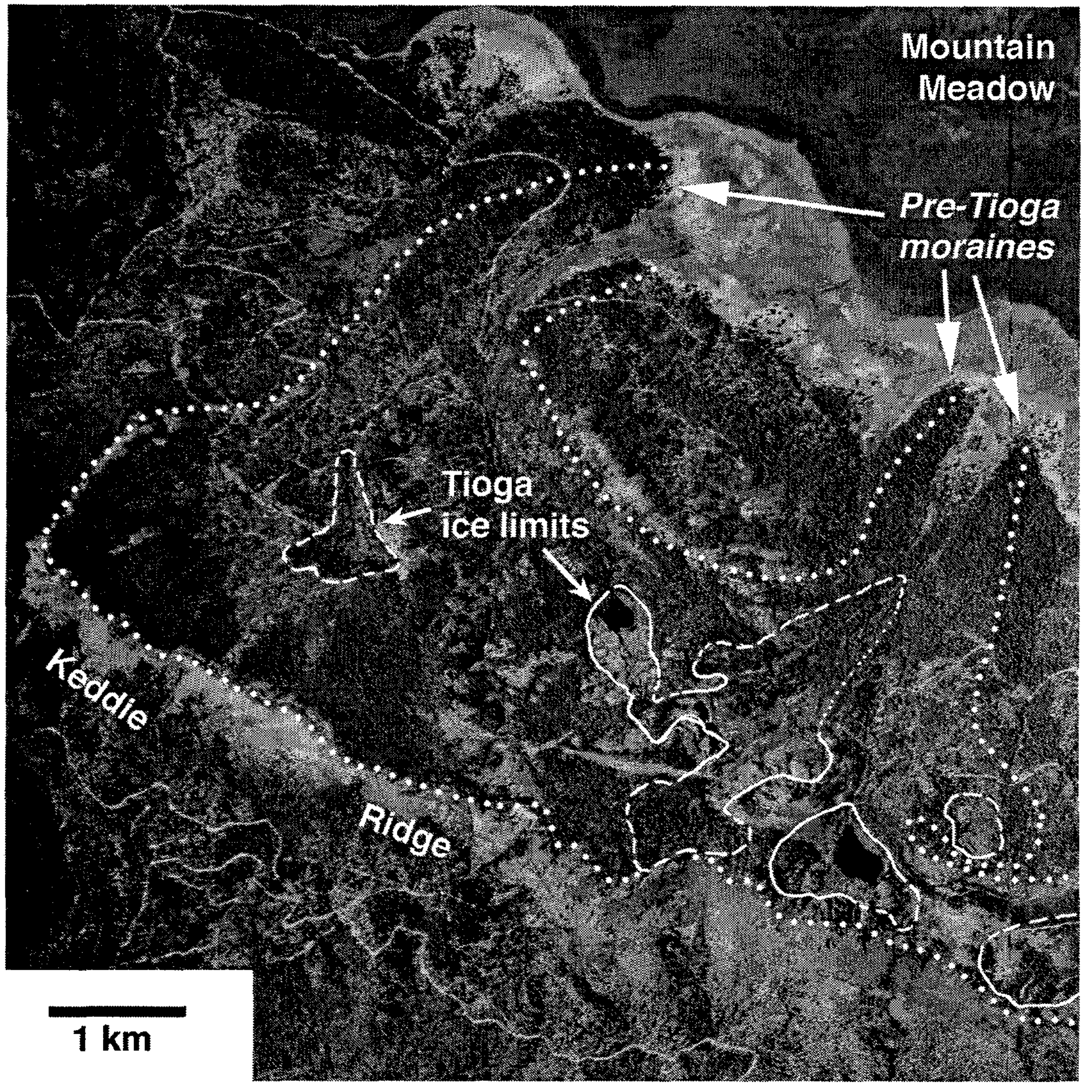


Figure 1.12: Vertical aerial photo of Keddie Ridge showing approximate limits of Tioga glaciers (thin lines) and pre-Tioga glaciers (bold dots). Glaciers flowed to the northeast from the ridge crest. Moraines of pre-Tioga glaciers are much more massive than those of the Tioga advance, crossing the range front and extending into Mountain Meadow. It is unclear whether Tioga or earlier glaciers occupied the steep southwest slope of Keddie Ridge.

Keddie Ridge Tioga glaciers was about 2-km long and flowed from ~2130 m to ~1650 m altitude at its maximum, although upper and lower ice limits are indistinct. Lateral moraines begin at about 1950-2010 m.

Two sets of large, morphologically distinct lateral moraines downslope from the Tioga moraines, which I term informally "Keddie Ridge" moraines, provide some of the strongest evidence for extensive pre-Tioga glaciation in the northern end of the range (Fig. 1.12). The moraines, which are volumetrically much larger than and extend well downstream from the Tioga moraines, begin at about 1700 m, cross the range-front, and terminate on the floor of Mountain Meadows at 1550 m; terminal moraines have been eroded or buried. The low slopes, broad crests, sparse boulders, highly oxidized soils, and extensive weathering rinds in clasts indicate that the Keddie Ridge moraines are substantially older than the Tioga moraines directly upvalley. The low altitude of the Keddie Ridge moraines suggests that the ELA there was much lower and ice extent much greater than during the Tioga advance.

Initial inspection of well developed weathering rinds (~1.0-1.5 mm, n = 6; Fig. 1.13) in fine-grained meta-andesitic cobbles from the B-horizon of the crest of the Keddie Ridge moraines indicates that they are much older than the Tioga moraines (< 0.1-0.2 mm rinds). This assessment is consistent with the better developed soils, fewer boulders, and morphologically mature slopes of the older moraines. According to Colman and Pierce (1981), andesitic cobbles with rinds of 1.0-1.5 mm are equivalent to those in pre-Tahoe till at Lassen Peak (Crandell, 1972), and Donner Lake Till at Truckee (Birkeland, 1964), although they could correspond to the Tahoe Till of Lassen Peak and the Hobart Till of Truckee within 1-σ error of the respective curves. In reexamining the weathering rinds of the pre-Tahoe and Donner Lake deposits, Colman and Pierce reasoned that they correlated with the penultimate glaciation (marine oxygen-isotope stage 6; Shackleton and Opdyke, 1973), although they did not have numerical ages to confirm this

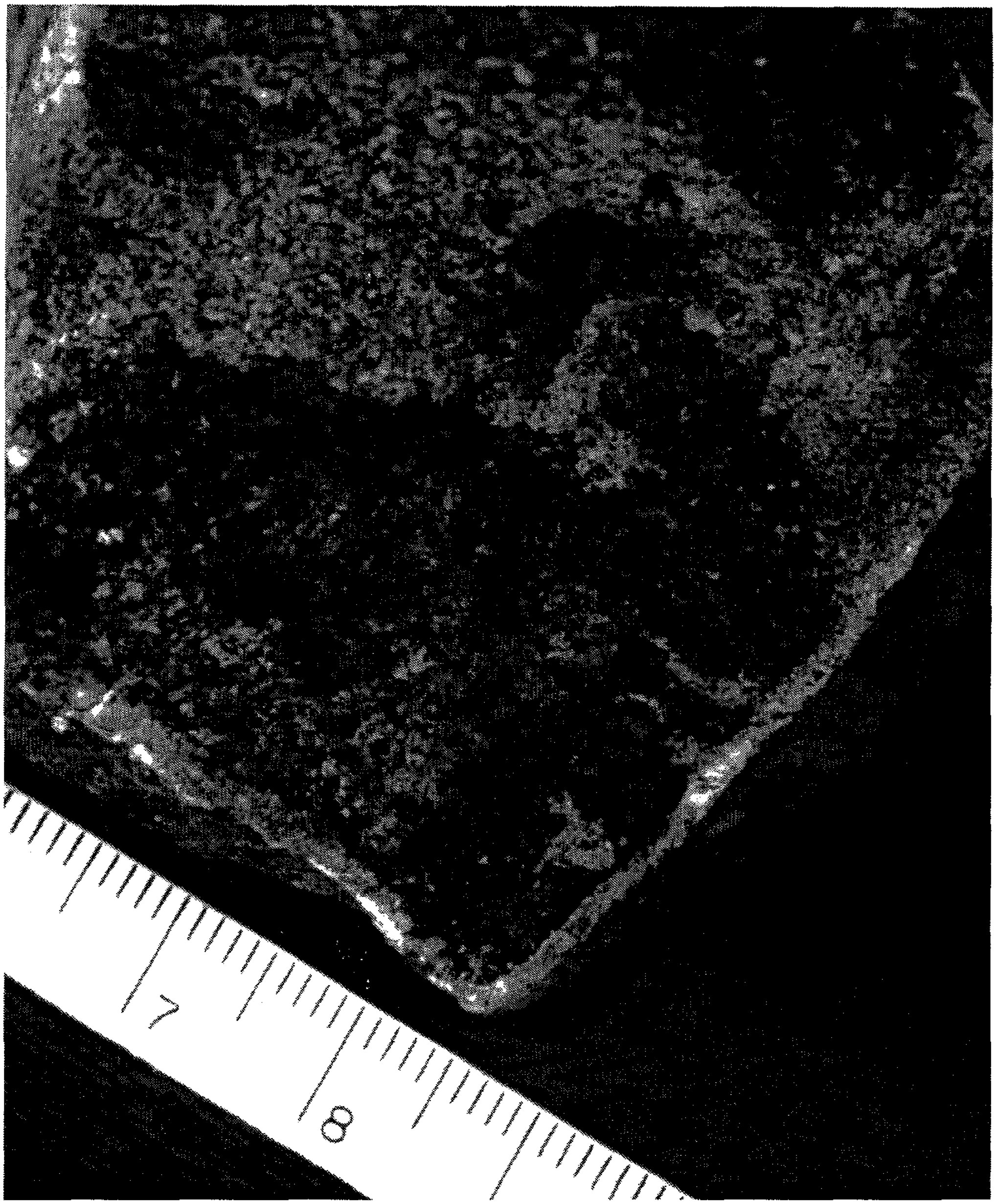


Figure 1.13: Close-up of 1-mm weathering rind developed in 5-cm diameter meta-andesite cobble from B-horizon of crest of northwesternmost pre-Tioga moraine, Keddie Ridge (see road crossing moraine, Figure 1.12). Five other clasts from the same site had weathering rinds 1-1.5 mm thick. Weathering rinds in clasts from Tioga moraines in the same drainage are all less than 0.1 mm thick. Units on ruler are in millimeters.

correlation. However, considering the order-of-magnitude difference in rind development of the Keddie Ridge moraines compared to the Tioga moraines, their assignment seems reasonable. It is notable that there are no distinct moraines between the Tioga and Keddie Ridge moraines.

The Keddie Ridge moraines also provide a limit on faulting in the area. The distinct linear break in topography and faceted slopes along the range-front, and the ponded sediments of Mountain Meadow, indicate that the range-front is fault-controlled (northeast-side down, normal); if so, the fault does not offset the moraines, and therefore the latest motion on the fault must predate formation of the moraines, presumably greater than ~140 ky BP. It is noteworthy that a fault which extends northward from the northwestern end of the Keddie Ridge range-front offsets Quaternary (?) basalt flows in a sense opposite to the range-front (i.e., down to the west). The opposite sense of throw and different ages of activity suggest that they are not the same fault.

Other Ranges

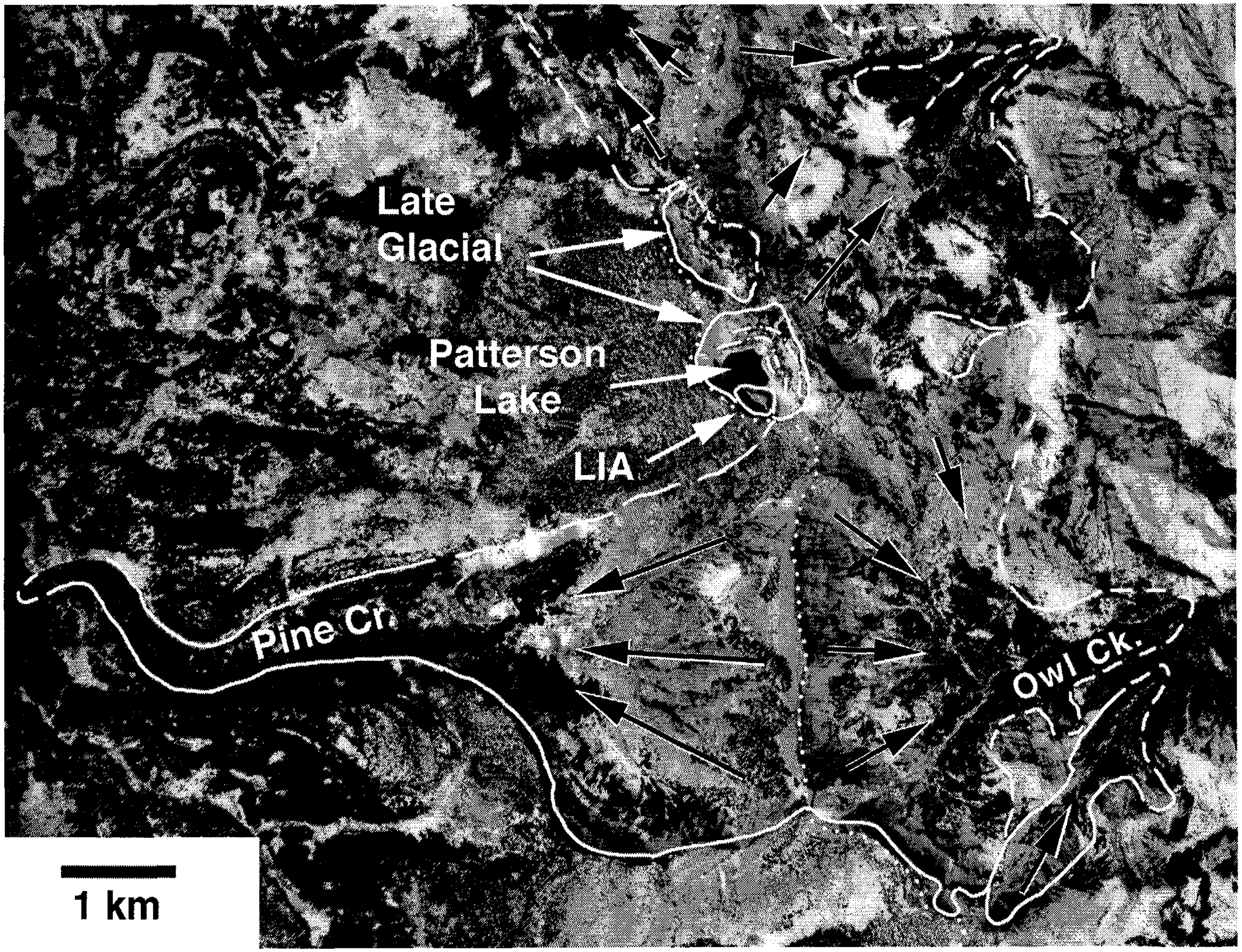
Although morphologically continuous with the Sierra Nevada, the Diamond Mountains (Fig. 1.2) are commonly considered a separate range. They encompass several northwest-trending ridges at the eastern edge of the Feather and Yuba river drainages, lying to the east of the northern end of the Sierra Nevada. Several of the peaks reach altitudes of 2150-2350 m, and supported small Tioga cirque glaciers, primarily on their northern slopes. Much of the area has been disturbed by extensive logging, however, and the volcanic bedrock apparently does not preserve glacial features well, making identification of extent of glaciation difficult. Moraines tend to be small and subtle in this area.

Evidence for late-Pleistocene and Holocene glaciation is strong in the southern half of the Warner Mountains, a north-trending, west-tilted mountain range in the northeastern corner of California (Fig. 1.2). The Warner Mountains reach altitudes near 3000 m, and supported late-Wisconsin valley glaciers on both its gentle western slope and its steep eastern slope. The volcaniclastic bedrock does not preserve striae or trimlines well, but well formed terminal, lateral, and recessional moraines, and dead-ice terrain along Pine Creek on the western slope (Fig. 1.14) indicate that it contained the largest late-Wisconsin glacier in the range, 6-7 km long, extending from an altitude of ~2750m to ~1950 m. Lateral moraines start at about 2300 m. Other less defined valley glaciers flowed off the west side of the ridge north of Eagle Peak (Fig. 1.2), but left only poorly defined moraines, making their size difficult to assess.

Nearly continuous, well formed cirques feeding off the crest to the east, many occupied by lakes and closed depressions, indicate that late-Wisconsin glaciers covered much of the uplands on that side of the crest in the south Warners, from Horse Mountain to Squaw Peak. Possibly the ice feeding the Pine Creek glacier was continuous across the crest with the glacier flowing eastward into Owl Creek. Moraines from late-Wisconsin maximum moraines are scarce on the steep slopes of the east side, but scattered moraine remnants and valley morphologies suggest that the glaciers there reached to altitudes at least as low as 2050 m (e.g., Owl Creek, Emerson Creek), and possibly lower. Local cirque glaciers formed on the east and north sides of some of the higher peaks in the north Warner mountains as well, but few appear to have flowed beyond the cirques.

The cirque on the northeast side of Warren Peak in the southern Warner Mountains contains two sets of moraines that postdate the late-Pleistocene maximum (Fig. 1.14). A low, rounded, vegetated moraine encloses and dams Patterson Lake, the largest glacial lake in the range. This moraine is well upstream of the maximum extent of late-Wisconsin ice in the drainage, but is weathered similarly to those deposits. In contrast, the cirque headwall above Patterson Lake contains a sparsely vegetated, unstable moraine that encloses a small basin. The upper half of the outside front of this

Figure 1.14: Vertical aerial photo showing approximate Tioga ice limits in Pine Creek and adjoining drainages, south Warner Mountains, California. Black arrows indicate direction of ice flow. Small moraines in northeast-facing cirques near Patterson Lake are from late-glacial (?) and Little Ice Age (LIA) advances. Dotted line shows crest of Warner Mountains. North is up.


Late
Glacial
Patterson
Lake
LIA
Pine Cr.
Owl Ck.
1 km

moraine is steeper than the angle of repose, suggesting that it may be ice-cored, still slowly flowing, and therefore young (i.e., Little Ice Age). Supporting this contention, I observed, over the course of several hours, a nearly constant fall of rocks from the headwall into the depression behind the moraine. Assuming this observation is representative, the bare bedrock exposed in the depression would very likely be covered within a century. It therefore seems likely that the glacier that formed this moraine has only recently disappeared.

DISCUSSION

The mapping I present above documents that substantial bodies of glacier ice existed in the northern Sierra Nevada north of the Stanislaus River during late-Wisconsin time, and probably during earlier glaciations. Although this portion of the range does not attain the high altitudes of the central or southern Sierra Nevada, latitudinal changes in temperature and precipitation during the last glaciation were sufficient to support a variety of ice fields, valley glaciers, and cirque glaciers. Thus, average altitudes between about 2400-2700 m on the plateau at the headwaters of the Mokelumne and Carson drainages were high enough to form an extensive ice field that covered all but the highest peaks in this portion of the range. At the northern end of the Sierran crest, the broad ridge separating the West Branch and North Fork of the Feather River supported a similar ice field, despite having average altitudes of only ~1850 m. This northerly decrease in the altitudes needed to form and maintain late-Pleistocene glaciers, about 500 m over roughly 200 km, mimics modern climate patterns of increased precipitation and decreased temperatures along the same profile (Dale, 1966).

ELA estimates of Tioga ice fields on the crest, from the highest occurrence of lateral moraines, also show a similar trend of progressively lower ELAs to the north (~500 m over the ~200 km distance; Fig. 1.15). This ELA gradient, ~2.5 m/km, is very

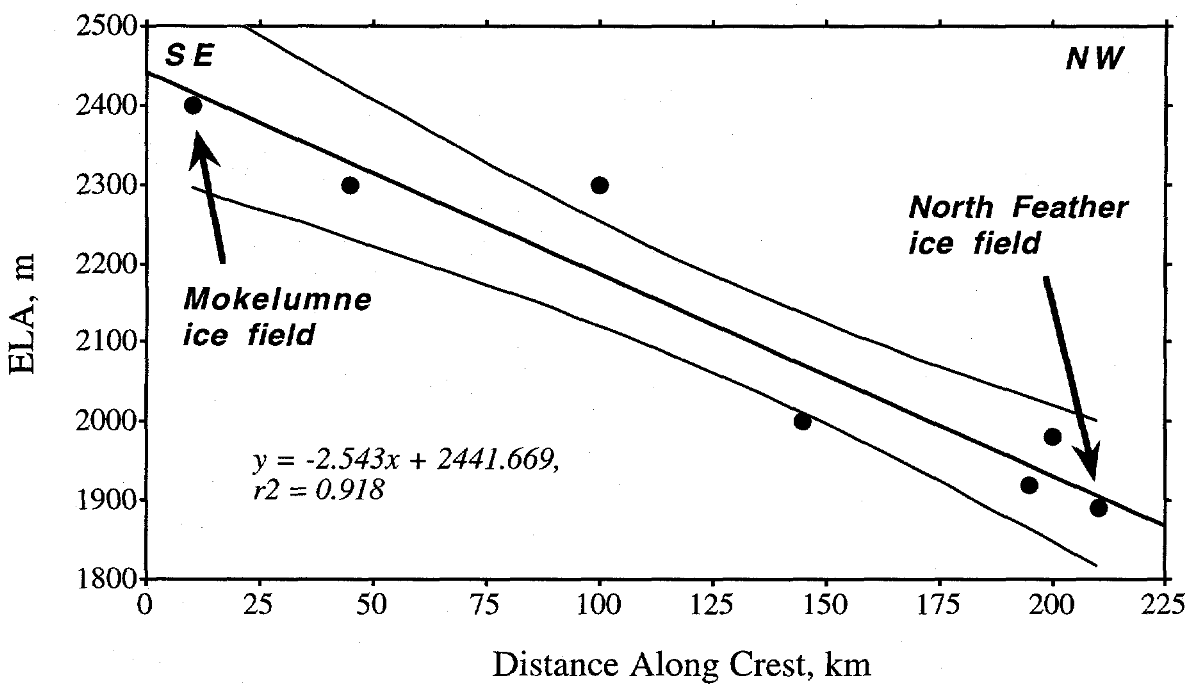


Figure 1.15: Profile of Tioga ELAs vs. distance along crest in the northern Sierra Nevada, from the Mokelumne ice field to the North Feather ice field at the northern end of the range. ELA estimates based on highest occurrences of lateral moraines in major ice fields along the crest of the range. Line is linear regression of ELA data (slope = -2.5 m/km to the NW), showing 2σ confidence interval (red curves) for slope and position. Maximum projection of data into profile is less than 3 km.

similar to that found by Burbank (~2 m/km; Burbank, 1991, Fig. 3) for the Tioga stade in the southern Sierra Nevada based on cirque-floor altitudes, and by Gillespie (Gillespie, 1991, Table 1) for early Tioga (3.1 ± 0.2 m/km) and late Tioga (2.2 ± 1.0 m/km) advances in the eastern Sierra Nevada using accumulation-area ratios. Furthermore, my ELA estimates overlap and match relatively closely with the northern end of Gillespie's profile, helping validate my estimates as well as extending Gillespie's profile to the northern end of the range. According to our combined data, Tioga ELAs along the crest decreased steadily from ~3400 m at the southern end of glaciation in the Sierra Nevada, to ~1900 m at its northern extreme. This represents a drop of 1500 m over 4° of latitude, from ~36°N to ~40°N. These numbers are remarkably close to those estimated for the Tahoe advance by Wahrhaftig and Birman (Wahrhaftig and Birman, 1965), although the decline in ELA along the crest is about 500 m greater than that estimated for the late-Pleistocene maximum by Porter (*in* Broecker and Denton, 1989, Fig. 6) in a generallized snowline profile of the American Cordillera.

ELA estimates of Tioga glaciers in a profile perpendicular to the crest at the northern end of the Sierra Nevada show a steeper gradient than the crest profile (~5.8 m/km; Fig. 1.16). This northeast-rising snowline gradient is about half that found by Burbank (13 m/km; Burbank, 1991, Fig. 4) from late-Pleistocene cirque floors in the southern Sierra Nevada. The discrepancy may in part reflect my necessary reliance on small cirque glaciers for the eastern portion of the northern Sierra Nevada profile, because larger glaciers do not exist in the Diamond Mountains east of the main crest. Such cirque glaciers often reflect local orographic ELAs and not the regional ELAs as do ice fields (e.g., Clark *et al.*, 1994a), and can therefore occur at much lower altitudes. Because the western control points are from ice fields, the cross-range gradient in Figure 1.16 is probably a minimum. Another more important reason for the discrepancy is that the ELA gradients reflect, to a large degree, the regional topographic gradient, due to the

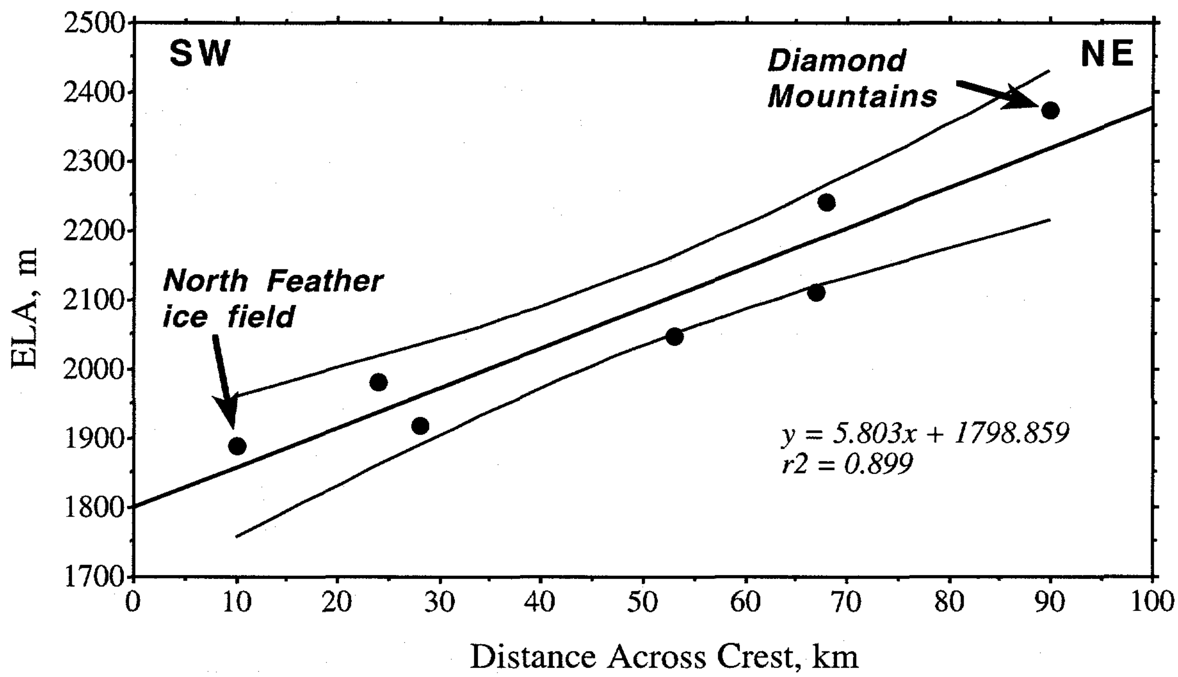


Figure 1.16: SW-NE profile of Tioga ELAs vs. distance across northern end of the Sierra Nevada. ELA estimates based on highest occurrence of lateral moraines from ice fields and cirque glaciers. Line is linear regression of ELA data (slope = 5.8 m/km to NE), showing 2σ confidence interval (red curves) for slope and position. Maximum projection of data into profile less than 5 km.

temperature/precipitation effects of orographic lifting (e.g., Porter, 1977, p. 105; Ohmura *et al.*, 1992, Fig. 1b, d). The general topographic gradient across the range where Burbank constructed his profiles is substantially steeper than it is where at the northern end of the range where I constructed my gradient; both ELA gradients mimic the regional topographic trends.

The overall picture of Tioga ELA trends in the Sierra Nevada indicates that they represent a range-wide change from modern climate, with ELAs lowered uniformly by ~700-900 m relative to Holocene conditions (Wahrhaftig and Birman, 1965; Burbank, 1991; Gillespie, 1991). The longitudinal and transverse ELA gradients resemble the regional patterns of modern temperature and precipitation in the range, in which summer temperatures decrease with altitude and, to a lesser degree, with latitude, and snowfall increases with altitude to maxima between ~2000 m (in the north) and ~3000 m (in the south) on the western slope (Dale, 1966). As with temperature, precipitation varies with latitude, but less than with altitude. Unlike the southern and central Sierra Nevada, late-Wisconsin ELAs cannot be directly compared to modern ELAs because the peaks are not high enough to support modern glaciers; the relative ELA depression between the two, ΔELA, must therefore be inferred. However, because the position and gradient of ELAs along the northern Sierra Nevada crest match those to the south (Burbank, 1991; Gillespie, 1991), and the modern temperature/precipitation gradients are relatively consistent along the same trend, it seems reasonable that the ΔELA between Tioga and modern conditions also remaines relatively constant. This inference is consistent with results from general circulation models, which suggest cooler and wetter conditions in California during full-glacial conditions as a result of a general intensification and southward shift of the Pacific storm track (CLIMAP Project Members, 1981; COHMAP Project Members, 1988).

Evidence for pre-Tioga glacier advances, in the form of moraines lying outside the Tioga maximum moraines, is scattered in the northern Sierra Nevada. The best-developed moraines typically formed from ice flowing out of the large ice fields along the crest, especially in those areas with crystalline bedrock. Good examples occur at the margins of the former Mokelumne, American/Truckee, Yuba, and Sierra Buttes ice fields. The Spanish Peak and North Fork Feather ice fields also probably preserve moraines from earlier glaciations. These older moraines, though perhaps not all correlative, indicate that, as in the southern half of the Sierra Nevada (Gillespie, 1982; Gillespie and Molnar, 1995), pre-Tioga glaciers in the northern Sierra Nevada were only marginally larger than those of Tioga age.

In contrast to the ice fields, individual cirques and valleys in the northern end of the Sierra Nevada that were glacierized in Tioga time often have no obvious deposits from earlier glaciations. Poor preservations of moraines formed on steep slopes or of poorly indurated and easily eroded lithologies, and cover by thick forest may all contribute to this apparent inconsistency. Notable exceptions, however, include sets of large, pre-Tioga moraines on the slopes of Mt. Ingalls and Keddie Ridge. The thick weathering rinds in clasts from the pre-Tioga "Keddie Ridge" moraines imply that they formed during or before marine oxygen-isotope stage 6. The absence of any obvious moraines separating the "Keddie Ridge" moraines from their Tioga equivalents suggests that any early-Wisconsin glaciers on the ridge were no larger or smaller than those of the late-Wisconsin advance. Detailed work is needed to assess this hypothesis.

CONCLUSIONS TO CHAPTER 1

For more than a century, geologists have noted the extensive evidence for large late-Pleistocene ice fields and glaciers that shaped the southern and central parts of the Sierra Nevada. The mapping and observations presented here, combined with those of earlier workers (Birkeland, 1964; Mathieson, 1981; Durrell, 1987), demonstrate that several unrecognized large ice fields and many smaller valley and cirque glaciers occupied the Sierran crest and adjoining high peaks during late-Pleistocene time.

The ice-field system along the Sierran crest was nearly continuous from south of the Mokelumne drainage to the headwaters of the Middle Fork Feather River, with two small interruptions, at the headwaters of the South Fork American and North Yuba Rivers. These small gaps may not have existed during earlier, more-extensive glaciations. However, a gap of nearly 40 km separated the northernmost two Sierran ice fields from those south of the Middle Fork Feather River. The ice fields, which probably reached their maximum dimensions in the Tuolumne drainage, remained large as far as the North Fork American River, then became progressively smaller as plateaus along the crest needed to maintain ice fields decreased in altitude and size. The effectiveness of such plateaus in forming large ice fields is best demonstrated in the headwaters of the Mokelumne drainage, where despite having few high peaks, a broad region that extended above the Tioga ELA supported one of the largest and most continuous ice fields in the range.

East of the crest, along the Carson Range, Diamond Mountains, and other small ridges, and at the north end of the Sierran crest, small isolated late-Wisconsin glaciers occupied north-facing slopes of the highest peaks. Most of these glaciers were no more than 1-3 km long and probably existed below the regional ELA, owing their existence to enhanced shading and snow-drift accumulation (e.g., Clark *et al.*, 1994a).

Estimates of ELA based on the highest occurrences of lateral moraines indicate that the Tioga snowline along the crest in the northern Sierra Nevada dropped about 2.5 m/km to the northwest, and rose by more than 5.5 m/km across the range to the northeast. These numbers agree well with those from the southern half of the range (Burbank, 1991; Gillespie, 1991). Thus, the climate associated with the last glacial maximum in the Sierra

Nevada appears to have resulted from an uniform change from the modern regional climate pattern along the entire range, rather than changing variably. This finding is consistent with the generalization that snowlines were lowered equivalently along most of the American cordillera during the late-Pleistocene glacial maximum (e.g., Porter *in* Broecker and Denton, 1989, Fig. 6).

Deeply weathered moraines outside Tioga ice limits indicate that the northern end of the Sierra Nevada was glaciated prior to late-Pleistocene time. These moraines are most common at the edges of former ice fields, but are also associated with sites of some smaller glaciers. Weathering rinds of cobbles from till indicate that large moraines downstream from the Tioga moraines at Keddie Ridge date from the marine isotope stage 6 or earlier. There are no obvious moraines between these two sets. These findings, though preliminary, suggest that early-Wisconsin glacier advances in the northernmost Sierra Nevada, in contrast with those in the southern Sierra Nevada (Gillespie, 1982; Gillespie, 1991) and at Lassen Peak (Gerstel, 1989), may not have been as extensive as those of stage 6 or stage 2. This inference clearly requires better age control on the Keddie Ridge moraines and other moraines in the region to be fully assessed.

CHAPTER 2: DEBRIS-COVERED GLACIERS IN THE SIERRA NEVADA, AND THEIR IMPLICATIONS FOR SNOWLINE RECONSTRUCTIONS

INTRODUCTION TO CHAPTER 2

Rock glaciers and ice-cored moraines have been studied extensively in mountains around the world, yet their origin and internal structure remain controversial. This situation reflects the remote setting of most of these landforms and the difficulties in observing their internal composition and flow characteristics. In this chapter I discuss my initial findings on the occurrence, development, and nature of rock glaciers and ice-cored moraines in the Sierra Nevada.

Wahrhaftig and Cox (1959) and White (1976) provide comprehensive reviews of early studies of rock glaciers. White also summarizes the debate on whether they develop from glacial or periglacial processes. Wahrhaftig and Cox characterized active rock glaciers in Alaska as permafrost phenomena, consisting of unsorted talus or moraine deposits mobilized by deformation of interstitial ice. Several subsequent studies support the view that all rock glaciers develop in this manner and are wholly non-glacial (e.g., Haeberli, 1985; Barsch, 1988). However, other studies (e.g., Foster and Holmes, 1965; Outcalt and Benedict, 1965; Potter, 1972; Benedict, 1973; Calkin *et al.*, 1987) demonstrate that some rock glaciers are cored by glacier ice mantled with a continuous and relatively thin debris cover. Arguments between the proponents of these two contrasting concepts have been passionate since Østrem (1971) and Barsch (1971) first clashed publicly over the origin of ice within rock glaciers. I will discuss this controversy with regard to Sierran rock glaciers in greater detail below.

In the Sierra Nevada, the only study that focuses on rock glaciers is by Kesseli (1941), although Matthes (1948) provides the first descriptions of ice cores in the voluminous Little Ice Age moraines. Kesseli speculated that rock glaciers were deposited underneath now-vanished debris-rich glaciers (i.e., a form of ground till), whereas Matthes ascribed surficial debris on Little Ice Age moraines to upward transport of subglacial debris along shear zones near the termini. Whalley (1974a, p. 43) briefly described rock glaciers in the southern Sierra Nevada as examples of ice-cored moraines in which the surficial debris originates from rockfall onto the glacier and acts as an insulative blanket.

Below, I describe rock glaciers and Matthes (Little Ice Age) moraines in the Sierra Nevada, and discuss their formation, extent, and climatic significance. My evidence indicates that most valley-floor rock glaciers (i.e., tongue-shaped rock glaciers of Wahrhaftig and Cox, 1959) and many Matthes moraines in the Sierra Nevada are ice-cored and glacial in origin, in agreement with Whalley (1974a). The greatly reduced rates of ablation below the equilibrium line of such debris-covered glaciers allow their ablation areas to grow much larger and extend much lower than those of equivalent "bare-ice" glaciers. To test this concept for past glaciations, I reevaluate deposits of the Recess Peak advance (Birman, 1964) in the southern Sierra Nevada that imply an apparently anomalous regional equilibrium-line altitude (ELA) gradient. The results indicate that the anomalous deposits are probably remnants of former debris-covered glaciers, and that the apparent ELA variance can be reduced by adjusting the reconstructed ELAs up-valley to reflect accumulation/ablation area ratios typical of modern debris-covered glaciers. The remaining ELA variance for the deposits results mainly from local orographic influences on accumulation and ablation.

Nomenclature

Uncertainty about the origin and structure of rock glaciers has led to several different classification systems. In general, these classifications are based either on gross morphology (e.g., lobate, tongue-shaped, or spatulate; Wahrhaftig and Cox, 1959, p. 389), internal structure (e.g., ice-cemented vs. ice-cored; Potter, 1972, p. 3027), or geomorphic position within a valley (e.g., valley-wall vs. valley-floor or cirque-floor; e.g., Outcalt and Benedict, 1965, p. 849). Although there may be some equivalency among these classifications (e.g., lobate = valley-wall = ice-cemented; C. Wahrhaftig, written communication, 1993), the important distinction here is whether the ice in rock glaciers is primarily glacial or periglacial (permafrost) in origin. As discussed below, my research indicates that valley-floor rock glaciers in the Sierra Nevada are glaciogenic (in contrast to valley-wall rock glaciers, which, *sensu stricto*, may be periglacial in origin). As such, they are part of a continuum of glacier types from bare-ice glaciers through debris-covered glaciers (or ice-cored rock glaciers) to debris-rich ice-cemented rock glaciers. The main difference between these glaciogenic types is the ratio of debris to ice. Some ice-cored rock glaciers may grade downstream into ice-cemented rock glaciers as the ice core melts and englacial debris is concentrated (C. Wahrhaftig, written communication, 1993). In view of their essentially glacial origin, these flowing masses are most accurately described as debris-covered glaciers, and it is in this sense that I use this term. However, I retain the general term "rock glacier," as originally used by Capps (1910, p. 360) and reemphasized by Potter (1972, p. 3027), to encompass all alpine debris-surfaced rock-ice flows, including debris-covered glaciers, without reference to internal structure or origin.

METHODS

Mapping

This study is based on mapping, field measurements, and aerial photo-interpretation of rock glaciers and moraines near the crest of the southern Sierra Nevada between latitudes 36°25' and 37°25' N, using 1:16,000-scale color and 1:60,000 and

1:80,000-scale black-and-white aerial photographs. I identify moraines and deposits of Matthes and Recess Peak advances on the basis of stratigraphic and topographic position, morphologic character, lichen cover, and general weathering characteristics. Matthes deposits are fresh, unstable, and generally lie next to modern glaciers. Matthes debris-covered glaciers are active, as shown by angle-of-repose or oversteepened terminal slopes. Recess Peak deposits typically lie 0.1-3 km downstream from the Matthes deposits and well upstream from the next lower (Tioga) moraines. Recess Peak deposits are little eroded, but have extensive lichen cover, some trees, few angle-of-repose slopes, and show incipient gullying of moraines.

Data resulting from the investigation were transferred to 1:24,000-scale USGS topographic maps. In addition, I recorded surficial details of some well developed rock glaciers in the southern Sierra Nevada on low-altitude, low-sun-angle oblique aerial photos.

Estimates of ELAs

Equilibrium lines, which separate areas of net accumulation and ablation on glaciers, are often used as proxies for alpine climatic snowlines, and thus climate, both modern and ancient. Two common methods of estimating ELAs of former glaciers (e.g., Meierding, 1982) use empirical ratios derived from modern bare-ice glaciers. The first ratio relates the former accumulation area to the total glacier area (accumulation-area ratio, or AAR), whereas the second ratio relates the altitude of the equilibrium line to the total altitudinal range of the glacier (terminus-to-headwall altitude ratio, or THAR)(e.g., Meierding, 1982; Leonard, 1984). The commonly used steady-state values for these ratios are AAR = 0.65 and THAR = 0.4-0.5. The accuracy of ELA estimates using these ratios depends on the former glaciers having mass balances and area/altitude distributions similar to those of modern bare-ice glaciers.

In order to assess past regional climate patterns in the southern Sierra Nevada, Burbank (1991) constructed ELA profiles both across and parallel to the crest of the range for late-Pleistocene and modern glaciers, using AAR and cirque-floor altitudes to estimate ELAs. From these profiles, he established that the ancient and modern gradients for each profile were approximately parallel (*ca.* 12 to 16 m/km for the cross-range profile, *ca.* 2 m/km for the crest-parallel profile). Burbank also constructed a crest-parallel ELA profile for the Recess Peak advance. He restricted these profiles to glaciers that were at least 300 m wide and had a northeasterly to northwesterly aspect to minimize ELA variability in cirques where local avalanching, drifting, or shading by cliffs (Meierding, 1982) cause unusually large accumulations of snow.

The Recess Peak deposits I consider in this chapter include those studied by Burbank (1991), which allows me to compare my ELA estimates with his. To improve this comparison, I apply the same method of calculating ELAs (AAR) and applied his "size-filtering" criteria, except where noted. I have also projected my ELA results onto approximately the same crest-parallel profile as Burbank. Because the deposits I studied all lie within 3 km of the transect, errors due to this projection are within the error of the methods of estimation.

OBSERVATIONS

Southfork Pass Rock Glacier

The active rock glacier that I studied most extensively in the field occupies a narrow, north-northeast-trending valley of South Fork Big Pine Creek directly below Southfork Pass (Fig. 2.1). The Southfork Pass rock glacier may be the longest in the Sierra Nevada. It extends 1.6 km from its head at 3740 m altitude to an active, angle-of-repose terminus at 3410 m and is 200-300 m wide (Fig. 2.2a, b). The debris-laden part of the rock glacier defines the ablation zone, approximately the lower 1.4 km (~90% by

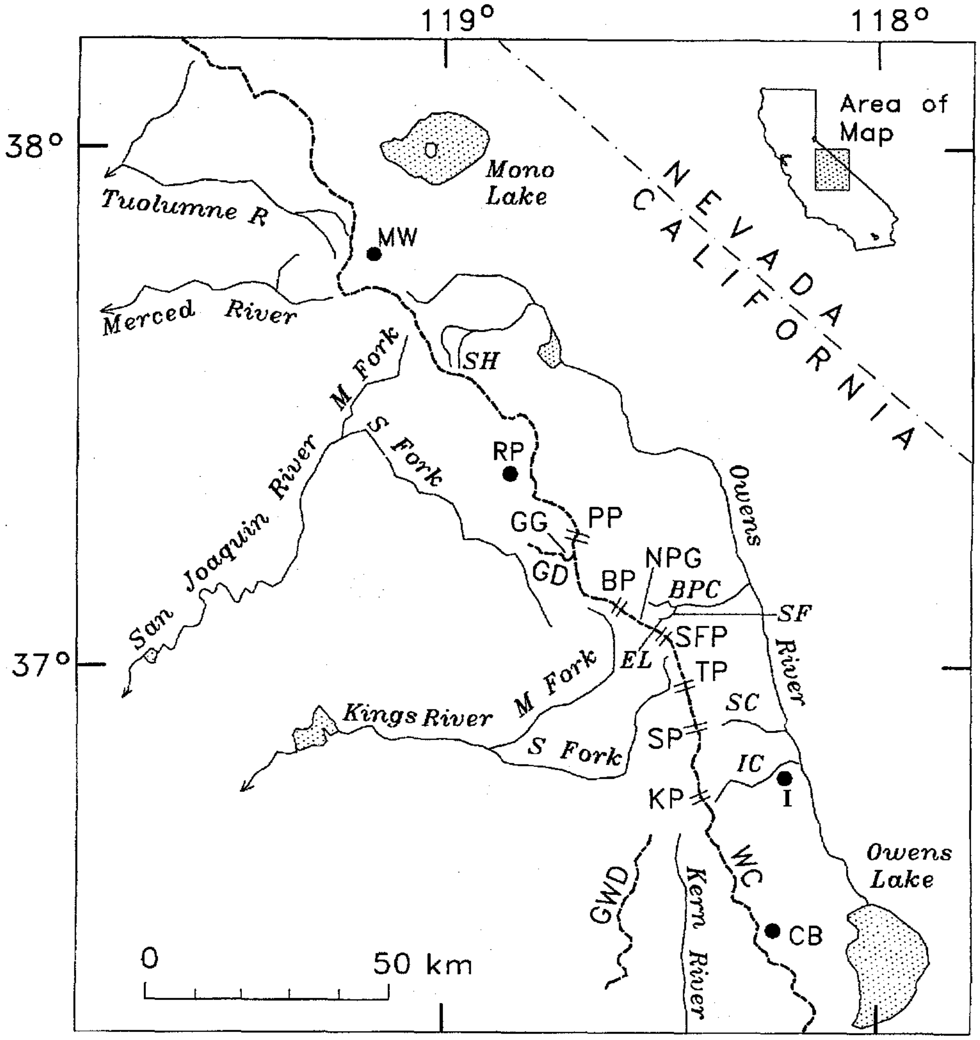
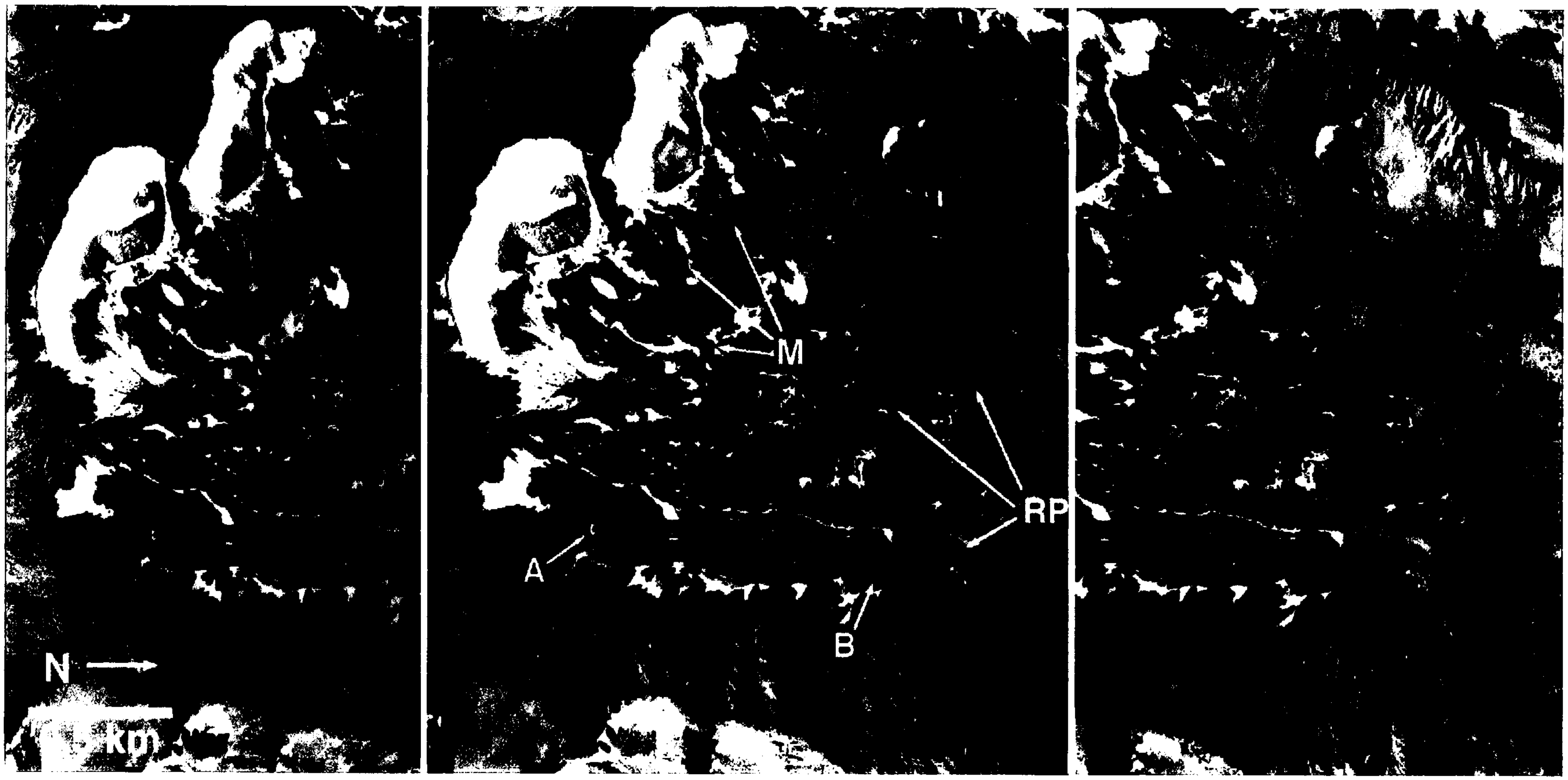


Fig. 2.1: Map of central and southern Sierra Nevada, California, showing major rivers and places named in the text. BP, Bishop Pass; BPC, Big Pine Creek; CB, Cottonwood Basin; EL, Elinore Lake; GWD, Great Western Divide; GD, Glacier Divide; GG, Goethe Glacier; I, Independence; IC, Independence Creek; KP, Kearsarge Pass; MW, Mt. Wood; NPG, North Palisade Glacier; PP, Piute Pass; RP, Recess Peak; SC, Sawmill Canyon; SF, South Fork; SFP, Southfork Pass; SH, Sherwin Creek; SP, Sawmill Pass; TP, Taboose Pass; WC, Whitney Crest.

Figure 2.2a. Stereotriptych of advancing Southfork Pass rock glacier (eastern third of photos). Photos show transverse ridges and longitudinal furrows, large thermokarst pond in upper half of rock glacier (A), and large boulder (B) near active, oversteepened terminus. Note large differences in total length and ratios of accumulation to ablation areas of rock glacier and adjacent (bare-ice) Middle Palisade Glacier, which is actively retreating from its Matthes-age end moraine (M). Ice limits of Recess Peak advance (RP) are shown downstream from each glacier. U.S. Forest Service aerial photos IN04 6-102, -103, 104, September 9, 1973.


N
M
RP
A
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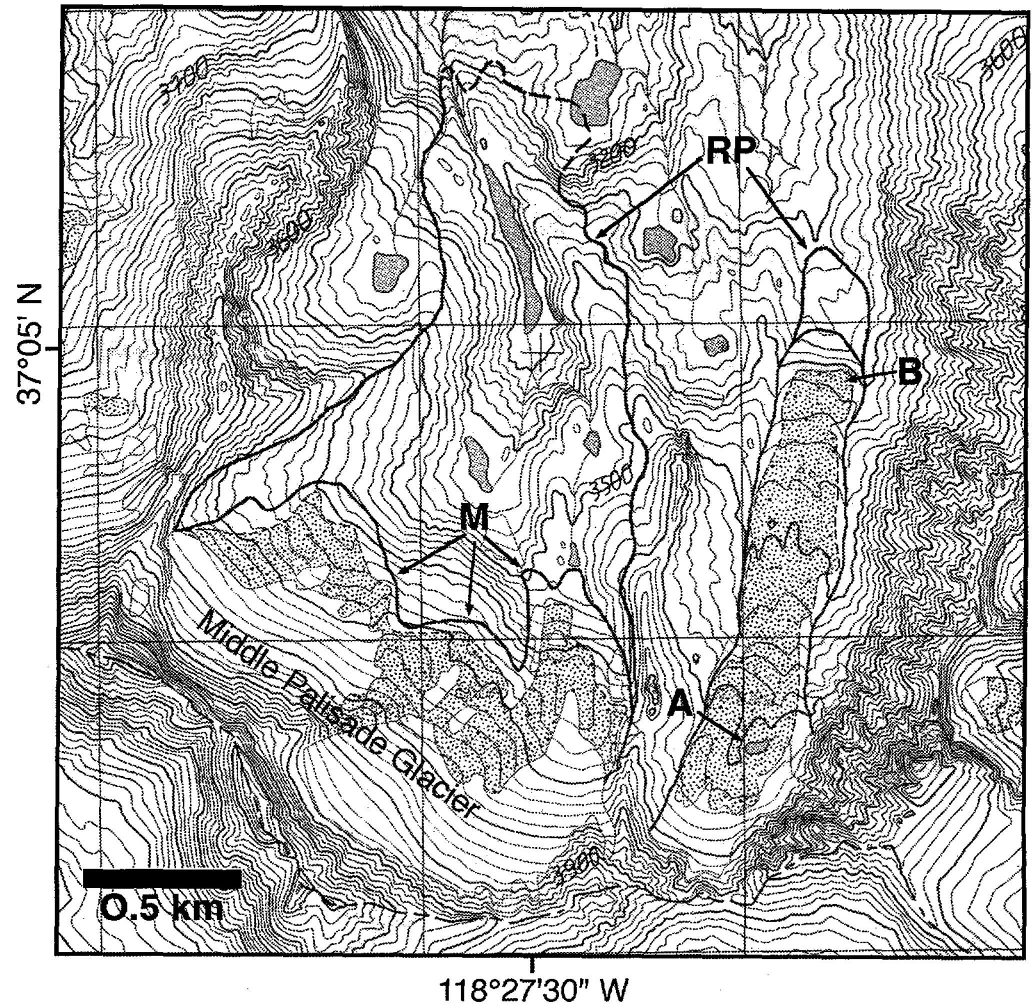


Fig. 2.2b. Topography of Southfork Pass rock glacier, showing same area as Fig. 2a. Moraine limits of Recess Peak (RP) and Matthes (M) advances are shown downstream from each glacier. Topography from USGS Split Mtn. 7.5' quadrangle, 20 m contours.

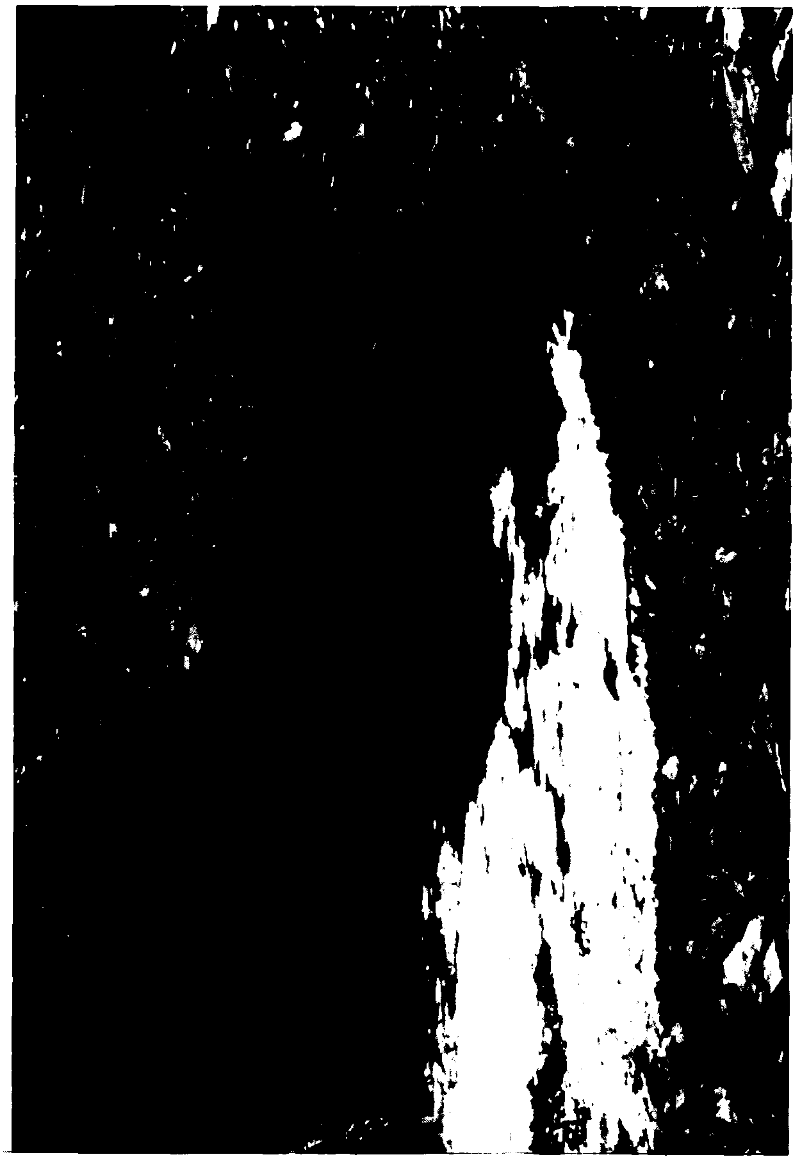
area). The uppermost 10% of the rock glacier, comprising the accumulation zone, is a steep ice-and-snow slope with scattered debris and a distinct bergschrund developed beneath the cirque headwall. The glacier ice/firn continues beneath the debris cover rather than terminating there. The transition along the axis of the glacier between ice and continuous surficial debris is distinct and coincides with an abrupt flattening of slope that forms a gently sloping bench on the uppermost section of the debris-covered surface. Similar benches, flats, or bowl-shaped depressions are common near the heads of other active valley-floor rock glaciers in the Sierra Nevada. The presence of higher lateral moraines at the edges of these benches and depressions indicates decreased snow accumulation and consequent lowering of the ice surface at the upper ends of these rock glaciers as climate warmed during the past century (e.g., Outcalt and Benedict, 1965).

The terminus of the Southfork Pass rock glacier is a broad, locally over-steepened slope ~70 m high that grades smoothly into vertically shorter, angle-of-repose lateral margins. A 1-m-deep pit excavated into the face of the terminus in late summer, approximately 10 m below the crest, revealed a saturated silty sandy matrix only slightly above 0°C. Silty water at ~0°C flowed (~0.1 m^3 min^{-1}) from a spring in the middle of the terminus about 15-20 m below the crest.

The debris-covered surface displays both transverse ridges (~1-2 m high) that are convex downslope from the margins and discontinuous meandering furrows that extend most of the length of the rock glacier near its central axis (Fig. 2.2). Running water can be heard along the furrows in the subsurface. The most distal furrow ends just short of the spring in the terminus. The surficial debris is coarse and angular, mostly 0.1 to >3 m in diameter, with local areas of silt and sand along the ridge crests and in the outer lateral and terminal slopes. The coarse debris is indistinguishable from the scree issuing from couloirs next to the eastern side of the rock glacier. This similarity, the absence of glacially abraded surfaces on the boulders, and the nearly continuous rockfalls onto the head of the rock glacier during the summer indicate that the surficial debris is derived primarily from rockfalls and rockfall avalanches at the head of the cirque (see also Whalley, 1974a, b). Distinct color boundaries in the surficial debris allow individual rockfall deposits to be traced to their sources. The source of rockfalls onto the glacier is highly fractured and unglaciated Tinemaha Granodiorite that comprises the headwall (Bateman, 1965); bedrock in the headwall above the adjoining Middle Palisade Glacier, composed mostly of Inconsolable Granodiorite, is more competent and delivers less debris onto that glacier. A continuous trough borders the eastern edge of the rock glacier from the headwall to the terminus and isolates the surface debris on the rock glacier from the sidewall talus (Fig. 2.2). This relationship requires that all the debris on the rock glacier came from the cirque headwall.

A small (< 25-m across in 1992), kidney-shaped thermokarst pond breaks the continuity of the debris cover about 1.2 km up-slope from the terminus (Fig. 2.2). The surface of the rock glacier next to the pond is morphologically indistinguishable from areas farther downslope. Vertical-to-overhanging walls along the entire southeastern half of the pond reveal clear, banded glacier ice beneath ~0.5 - 1 m of debris (Fig. 2.3). The exposed ice is free of the large boulders that blanket the surface, although it does contain several thin, dark, parallel silt layers, and one layer with pebble- to cobble-size debris, that dip up-glacier about 35 - 45°. The silt layers average about 0.5-1 cm thick, and are separated by zones of clear banded ice approximately 10-40 cm thick; the alternating translucent and milky bands within the clear ice are typically 5-10 cm thick and likely represent annual layering from snowfall (Outcalt and Benedict, 1965, Fig. 6). During the summer, rockfall onto the glacier above the pond is nearly constant but leaves only silt, sand, and small clasts in the accumulation zone. This observation explains the absence of boulders within the ice, and supports the inference that the greater momentum of larger

Figure 2.3. View to southeast of glacier-ice wall of thermokarst pond in Southfork Pass rock glacier. See text for description. Ice-wall is ~4 m tall. August 25, 1992.



clasts in a rockfall generally carries them beyond the steep slope of the accumulation zone (Potter, 1972).

Aerial photos show that the pond formed sometime between 1947 and 1962, and has continued to grow since then. This steady enlargement indicates that the pond is below the modern ELA, and that unless it is buried by a large rockfall or increased snowfall, or freezes in response to long-term cooling, the pond may eventually behead the rock glacier. The existence and growth of the pond also indicates that, at least in this area, the glacier ice is temperate (i.e., at the melting point). The absence of similar ponds farther downstream probably in part results from thicker debris cover there, a condition that would effectively prevent the local excess melting needed to form such ponds. Alternatively, such ponds may result from collapse of an internal drainage system within the ice core, as indicated on historical aerial photos of a pond that formed on the Powell rock glacier, South Fork San Joaquin River (S. Konrad, personal communication, 1995), and as also indicated by observations of a cavernous drainage tunnel within the ice cored moraine of the Goethe Glacier (Clark *et al.*, 1994b, p. 359).

The oversteepened upper slope on the terminus of Southfork Pass rock glacier and those of other rock glaciers throughout the range, obvious on aerial photos and in the field, are the primary evidence that they are active (Wahrhaftig and Cox, 1959, p.392). In addition, down-valley movement of large boulders on the Southfork Pass rock glacier (e.g., B, Fig. 2.2) demonstrates at least several meters of glacier movement between 1947 and 1981 (Fig. 2.4). An average value of about 0.4 m/yr is comparable to reported values for other similar ice-cored rock glaciers (e.g., 0.14 m/yr, Potter, 1972, p. 3049; ~0.3 m/yr, C. Wahrhaftig, written communication, 1993).

In combination, these observations indicate that Southfork Pass rock glacier is glacial in origin, is cored by glacier ice, and is therefore described more accurately as a debris-covered glacier. Evidence of contemporary permafrost in the area is absent. I

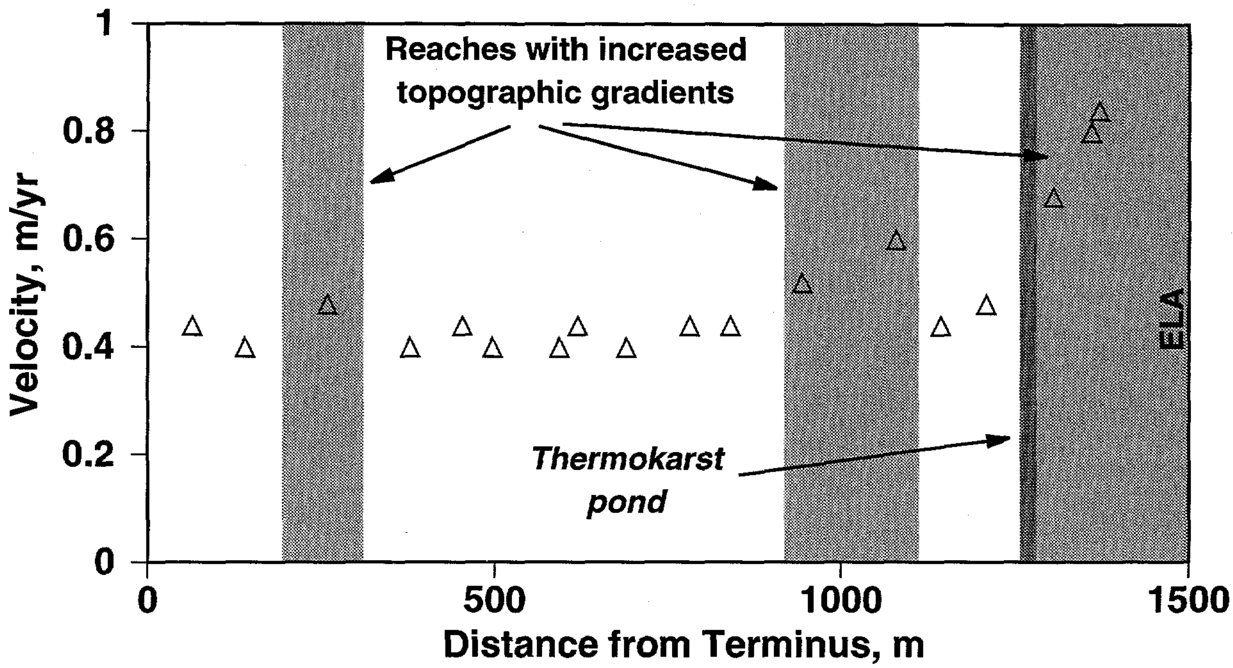


Figure 2.4. Modern down-glacier velocities from displacements for 18 large boulders along the axis of the Southfork Pass debris-covered glacier between 1947 and 1972, measured from enlargements (~1:8000) of USFS aerial photographs GS-CQ 6-142 (1947) and IN04 19-103 (1972). Velocities for the lower portion of the glacier are nearly constant at 0.4 m/yr. Measurement uncertainties are ±0.04-0.08 m/yr. Zones of high velocity (grey shading) coincide with increased slopes on glacier. Location of thermokarst pond shown by dark grey strip. The approximate estimate of 0.1-0.2 m/yr reported by us in our paper was for a boulder well off the axis of the glacier. Velocities rise toward the equilibrium line of the glacier.

emphasize that, although unusually large, this debris-covered glacier is morphologically, texturally, and structurally indistinguishable from most other active valley-floor rock glaciers in the Sierra Nevada. Such rock glaciers in the Sierra Nevada occur between 36°17' N and at least 38°10' N latitude. I infer that most of these valley-floor rock glaciers are actually debris-covered glaciers, and that the general characteristics described above for the Southfork Pass rock glacier can be reasonably applied to other Sierran rock glaciers.

These findings contradict the views of Haeberli (1985, p. 7) and Barsch (1988, p. 79), who insist that all rock glaciers are permafrost phenomena, primarily on the basis of mapping and geophysical measurements in the Swiss Alps. However, some of their inferences about rock-glacier development (e.g., Haeberli, 1985, Fig. 4; Barsch, 1988, Fig. 4.9) cannot be reconciled with my direct observations of rock glaciers in the Sierra Nevada. The blanket concept of a permafrost origin for all rock glaciers is thus untenable.

Valley-Wall Rock Glaciers

Valley-wall, or lobate, rock glaciers in the Sierra Nevada, in contrast to valley-floor rock glaciers, do not appear to be strictly glacial in origin. These flows are spatially and kinematically distinct in that they typically form on steep valley walls below unglaciated couloirs, and typically flow only short distances toward the valley axis. Such rock glaciers rarely flow beyond the steep slopes on which they originate, suggesting greater viscosities than valley-floor rock glaciers. Further distinctions include width-to-length ratios typically greater than 1.0, less complex and organized surface morphologies (i.e., less-distinct transverse ridges and absence of longitudinal furrows), and no active or remnant ice accumulation zone (characterized by a bowl-shaped depression at the heads of ice-cored rock glaciers). Also, valley-wall rock glaciers appear to be less sensitive to geographic aspect than valley-floor types; many occupy south-facing valley walls.

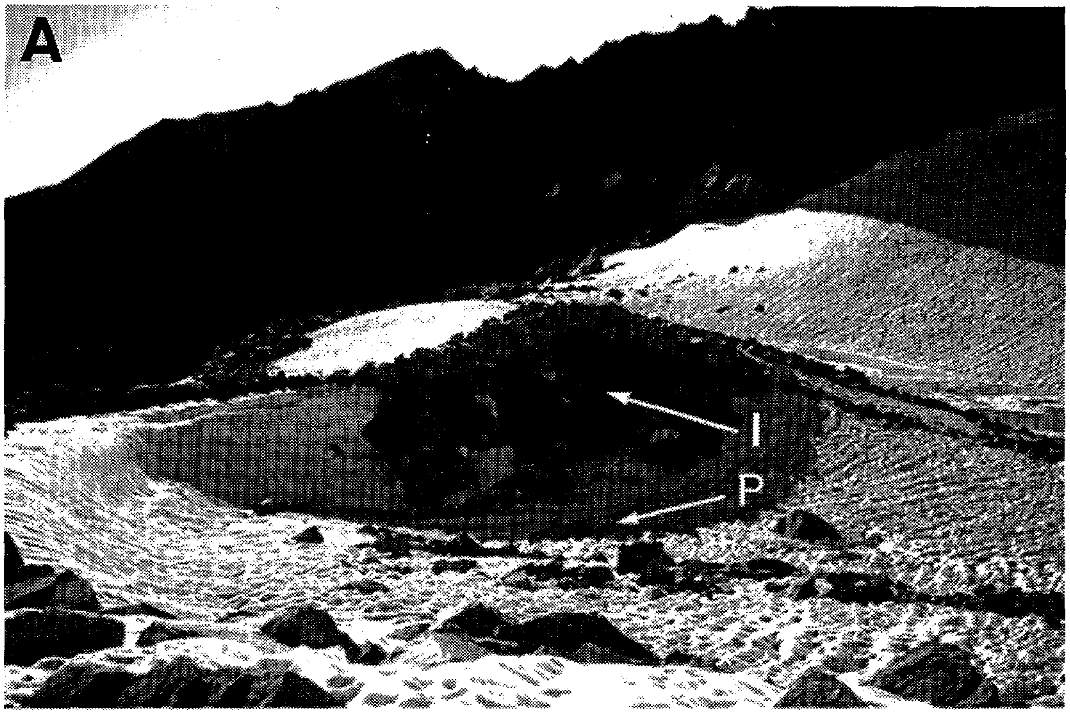
On the basis of these characteristics, valley-wall rock glaciers appear to be taluses that have mobilized due to internal shearing along ice lenses or ice concentrations within the rock debris (Wahrhaftig and Cox, 1959; White, 1976, p. 81). The absence of other modern permafrost features in the Sierra Nevada implies that most of the internal ice is primary and forms from rockfall-buried snow or winter/spring snow-and-rock avalanches (Outcalt and Benedict, 1965, p. 856), with only minor additions from refrozen precipitation (e.g., Barsch, 1992, p. 178). Thus, a progression of rock-glacier forms is likely, ranging from debris-covered glaciers to slightly mobilized taluses with only local internal ice. The forms more closely associated with true glacier ice appear to be more important to paleoclimatic studies, because the talus-derived deposits appear to form under substantially more variable climatic conditions (Calkin *et al.*, 1987, p.66). For example, in several of the drainages near the town of Independence (Fig. 2.1) on the southeastern slope of the Sierra Nevada, valley-wall rock glaciers reach altitudes more than 200 m below the lowest contemporaneous debris-covered glaciers.

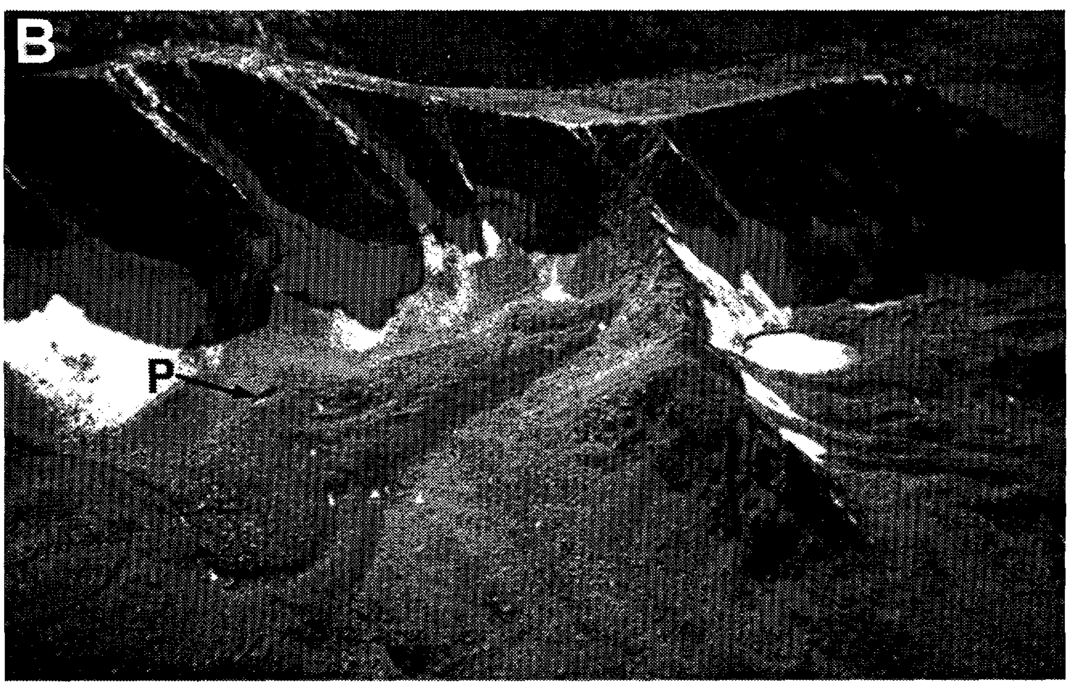
Ice-Cored Moraines

End and lateral moraines of Matthes age in the Sierra Nevada are typically quite different from older associated cirque deposits. Most Matthes moraines are voluminous and bear a strong resemblance to debris-covered glaciers in the range. These moraines have high, angle-of-repose (or, locally, steeper) outer slopes; consist of coarse, relatively unweathered rubble; and bear little or no vegetation and scattered lichens (Fig. 2.5).

Few if any Matthes moraines adjoin expanding glaciers. Many abut shrinking glaciers (whose surfaces have ablated to levels well below the crests of the moraines), some are dissociated from such glaciers, and for some, no glacier survives. Nevertheless,

Figure 2.5. Ponds along crests of Matthes-age moraines, Sierra Nevada. (A) View southeast showing ephemeral pond (P) that occupies crest of moraine of unnamed glacier (right) ~1 km southwest of Elinore Lake, South Fork Big Pine Creek. Wall of pond exposes clear glacier ice (I) beneath ~1 m of rubble. August 10, 1980. (B) Aerial photograph looking southwest over Goethe Glacier (center, left), "Wahoo" glacier (right), and associated moraines of Matthes-age on north side of Glacier Divide. Pond (P) occupies crest of moraine next to Goethe Glacier; larger ponds lie in depression between present "Wahoo" glacier and end moraines. Wall of pond next to Goethe Glacier reveals clear ice beneath 1 to 2 m of rubble. Glacier ice is exposed in moraine above ponds at "Wahoo" glacier. Matthes-age moraines of both these retreating glaciers have angle-of-repose faces that indicate current downslope movement. September 27, 1976.


A
I
P


B
P

the steep, unstable outer slopes of most of these moraines are most reasonably explained by recent or active internal flow, most likely related to an ice core. Indeed, Matthes (1948) reported ice cores in some of these moraines, exposed in both natural trenches and on downstream faces.

In addition, small (< 10 m diameter), steep-walled ponds occupy the crests of many Matthes moraines, further implying ice cores. M. Clark (personal communication, 1993) and I have inspected two such ponds on moraine crests at ~3680 m above Elinore Lake and above Goethe Lake, respectively 3 and 30 km northwest of the Southfork Pass rock glacier (Figs. 2.1 and 2.4). The ponds we inspected are very similar to the one on the Southfork Pass rock glacier: the steep walls are clear ice beneath about 1 m of debris cover. Although distinct annual layering is not apparent in the ice walls of the moraine ponds, a glacial origin provides the simplest and most reasonable explanation for the ice.

Location on moraine crests allows such ponds to drain easily as they grow, and the disappearance of some ponds can be seen on successive aerial photos. Periodic emptying may limit their size or life-span by allowing the debris cover to re-form, once a bottom lag deposit is exposed by draining.

Such ponds are firm evidence of ice cores. Because the moraines that I inspected are morphologically indistinguishable from most other Matthes moraines in the region, I conclude that other Matthes-age moraines of the Sierra Nevada that have angle-of-repose slopes also have cores of glacier ice. Furthermore, a morphologic progression links ice-cored moraines to debris-covered glaciers (Fig. 2.6). These moraines are evidently transitional between stagnant ice-cored moraines and active debris-covered glaciers.

DISCUSSION

The apparent widespread occurrence of cores of glacier ice in active valley-floor rock glaciers and Matthes moraines of the Sierra Nevada emphasizes that relatively thin

Figure 2.6. Oblique aerial photograph of ice-cored Matthes-age end-moraine complex adjacent to "modern" (1972) terminus of North Palisade Glacier, showing angle-of-repose distal slope that indicates actively flowing moraine margin. This moraine complex appears to be intermediate between a stable ice-cored moraine and an active debris-covered glacier (ice-cored rock glacier). Note coarse rockfall debris on upper ablation zone of North Palisade Glacier and adjacent unnamed glacier, and englacial debris exposed in lower part of these glaciers. View to southwest. Photo #72R2-151, August 24, 1972, USGS, Tacoma, WA.

blankets of surficial debris insulate and preserve the underlying ice. As a result, debris-covered glaciers can maintain a positive mass balance under adverse climatic conditions, and their ablation areas can expand beyond the size that will maintain bare-ice glaciers in a balanced state. This effect is demonstrated by comparing the modern glaciologic character of the Southfork Pass debris-covered glacier (presently advancing at ~3410 m) to that of adjacent Middle Palisade Glacier (presently thinning and retreating at ~3700 m, even though it originates at a higher altitude and has a much larger accumulation area; Fig. 2.2). Similar relationships between debris-covered glaciers and adjoining bare-ice glaciers are common in the southern Sierra Nevada.

Insulating Properties of Debris Cover

Several heat-transfer processes potentially operate within supraglacial debris during the summer, in particular conduction of ambient heat from the atmosphere, absorbed incident radiation, forced convection from wind, and nightly air turnover. Percolation of rain and snow melt introduce little heat to temperate glaciers (Paterson, 1981, p. 304), and presumably less to those covered with debris. Condensation and its effects are probably insignificant under continuous debris.

Depending on its thickness, debris cover either accelerates or retards ablation. In general, the effects of increased solar absorption by low-albedo debris is counteracted in thicker debris blankets by reduced heat conduction and decreased air movement through the debris layer. Furthermore, warmer, less-dense daytime air does not sink into a porous debris cover, but colder nighttime air will replace any warmer air beneath it in the debris. This diurnal process, a form of Balch ventilation (Thompson, 1962, p. 214), will tend to maintain the lower part of the debris cover at the minimum daily temperature. At the altitudes of Sierran rock glaciers, that temperature is close to freezing, even in summer (Dale, 1966).

Heat Transfer Through Debris During the Ablation Season

In this section, I briefly investigate the relative importance of the major processes, mentioned above, which transmit heat through a debris layer into the underlying ice during the ablation season.

In terms of ablative processes, debris-covered glaciers are distinct from clean glaciers only during the ablation season after snow from the previous winter has melted away to expose the regolith. In the higher elevations of the Sierra Nevada, to which modern rock glaciers are restricted, this exposure commonly does not occur until late June or early July, and typically lasts only until the first major snow storms in late September or early October. Thus, for this analysis, I define the average rock-glacier ablation season as the three-month period extending from the summer solstice to the fall equinox.

Daily heating: First I discuss the effect of daily heating on the surface at the midpoint of a typical Sierra Nevada debris-covered glacier during the ablation season. This average position can be represented by a moderately inclined (e.g., 5-10° N) surface at 3500 m and 37° N latitude in a northeast-facing cirque. There are two main sources of heat input: radiant heating by the sun and heat to or from the atmosphere. Radiant heating of the glacier surface (debris or ice) is determined by the sun altitude and azimuth over the course of a day (and shading from local cliffs), the slope and azimuth of the glacier itself, and the atmospheric conditions (especially cloudiness). Air temperature is an important factor in the transport of sensible heat by conduction to or from the exposed debris, and also in the transport of heat across void boundaries within the layer, in the case of loosely packed debris.

Daily heating of the surface due to direct radiation is important only for debris layers that are thinner than the diurnal heating front can penetrate. Schieldge *et al.* (1982) demonstrate that diurnal temperature fluctuations in soils become negligible at depths of

~0.5 m. Simple calculations of the daily heating in coarse granitic regolith show similar damping of the daily heating cycle in the surficial cover of rock glaciers. In application, these results mean that rocks or ice buried below about a half meter of debris (i.e., most of the ablation areas of Sierran rock glaciers) will not "feel" the effects of daily solar heating.

This finding enables one to put aside effects due to radiant heating and to focus on seasonal temperature profiles based on average air temperatures. Direct and continuous meteorological measurements in the Sierra Nevada are sparse, but air temperature conditions at the hypothetical glacier can be estimated by extrapolating from measurements at Grant Grove (Dale, 1966). This location, at ~46°45' N and 2000 m elevation, is relatively close to the "glacier" and is approximately upwind on the west side of the range. Applying an average wet adiabatic lapse rate of 6.5°C km^{-1} (Powell and Klieforth, 1981, p. 12) indicates that the average minimum and maximum temperatures at the elevation of the glacier during July, August, and September are 0° and 14°, -2° and 14°, and -3° and 12°C, respectively. In the simplest form, one can calculate the heat flux through the layer under steady-state conditions, with the surface temperature held fixed at the average summer air temperature (about 6°C, from above) and the basal layer at 0°C (the temperature of temperate glacier ice). The heat flux, f_Z, through the layer at depth z is: $f_z = -K\,\partial T/\partial z$ (Carslaw and Jaeger, 1980) where K is the thermal conductivity and T is temperature. For the simple case here, I assume representative values of $K = 0.003$ cal cm^{-1} sec^{-1} °C^{-1} (the value for gravel)(Kahle, 1980, p. 264), $DT =$ 6°C, and $Dz = 200$ cm; thus $f_{200} = -9 \cdot 10^{-5}$ cal cm^{-2} sec^{-1}.

This value of f_Z transmits about 715 cal cm^{-2} to the ice during the ablation season, or enough heat to melt 8.9 g (~10 cm thickness) of ice already at 0°C. For the extreme case, increasing K by a factor of about 2 (typical of solid granite) roughly doubles the amount of ice melted to 18 g (~20 cm thickness of ice) per summer. Similarly, melting would double if the debris layer is thinned to 100 cm or if the average summer surface temperature of the layer reached 12°C as a result of radiant heating. This model overestimates the conductive heat flux through the layer, because actual temperatures below the penetration depth of the diurnal heat wave(~0.5 m) will not fully equilibrate with the average summer temperature. The gradient will instead approach freezing asymptotically with depth. The high ablation rates associated with very thin cover (<4 cm) produce local instabilities: rapid ablation under such debris will tend to add new debris to the surface, from the ice below or the glacier upstream. As a result, the glacier will tend to "heal" areas of thin cover and rapid ablation.

One can compare the numbers established above with those estimated for clean ice, for which incident radiation becomes quite important. Extrapolating to 3500 m from low-elevation, clear-day observations of Berdahl *et al.* (1978), maximum total daily incident radiation on a horizontal surface at the idealized rock glacier varies from about 34.2 to 24.5 MJ m^{-2} day^{-1} between the summer solstice and the fall equinox. For clean, multiyear ice with an albedo of about 0.5 (Paterson, 1981) this radiation would introduce a maximum (because the estimate assumes no clouds) of 17.1 to 12.2 MJ m^{-2} day^{-1}, or 400 to 300 cal cm^{-2} day^{-1}, to the ice. This converts to roughly 30 kcal cm^{-2} per summer, equivalent to melting about 4 m of ice. For ice with a continuous but very thin cover of fine, leucocratic debris, albedo will be reduced to that of the debris (~0.15; Abrams and Brown, 1984, p. 4-18), with consequent heating increased to ~55 kcal cm^{-2} per summer, enough to melt almost 7 m of ice. These estimates, although not accounting for second-order ablation effects such as wind, condensation, cloudiness, rain, or evaporation, basically agree with observational studies (Østrem, 1959; Whalley, 1974a; Loomis, 1970; Lundstrom *et al.*, 1993).

Several other processes will tend to lessen the radiant heating of and heat conduction through a debris layer:

Nocturnal air inversion: Nightly overturn of air should decrease the penetration of the diurnal and seasonal heating fronts into the debris cover. As surface temperatures drop at night, an unstable density gradient is created between overlying cold surface air and the lighter air within the upper layer of debris cover, still warm from the downward conduction of midday heat in the debris. The intralayer air will tend to rise and be replaced by denser surface air. Density differences at the chosen elevation (3500 m, or 500 mm Hg) can be estimated from the average observed temperatures shown above. If air within the layer that is warmed during the day is held to be at roughly 5°C with a dew point at 0°C, and nighttime air averages -2°C with a dew point at -5°C, the former will have a density of 0.8324 g/l and the latter will have a density of 0.8612 g/l (Weast, 1976, p. F-9). This density shift, roughly 3%, sets up nocturnal cooling, in which air warmed by the radiantly heated debris is continually replaced by cooler, denser air from above. Even saturated air at 0°C, such as occurs deeper within the debris layer, will have a density less than that of surface air that is below freezing. Conversely, heating of the surficial air during the daytime sets up a stable density gradient such that heat is introduced at depth only through conduction, and not by convection. In effect, the debris layer acts as a convective cooling ratchet: the entire layer is cooled convectively each night toward the nightly minimum temperature, but is not heated convectively at all during the day. In a practical sense, the nightly convective heat loss will only be effective in the coarser-grained upper portion of the debris layer, because the frictional resistance to air circulation within the gravel-sand-silt mixture is unlikely to be overcome by the small density differences cited above. However, surface winds will also not significantly stir the air below the near-surface debris, thus preventing penetration of sensible heat into the layer by forced convection.

Evaporation of rain and snow-melt: Evaporation likely plays a substantial role in decreasing the amount of heat that penetrates the debris cover on glaciers. In the Sierra

Nevada, summer air is typically dry, except during the occasional influx of summer moisture from the south. In contrast, deeper portions of the regolith on debris-covered glaciers are saturated from snow- or ice-melt, with films of water coating the clasts. Evaporation of this water into the surface air will help maintain cool temperatures in the overlying rocks. Similarly, occasional afternoon thundershowers will block the sun and cover the surface with relatively cool water. Such rainstorms often occur in mid afternoon, when surface temperatures otherwise are greatest. Sensible cooling and evaporation of the rain water should effectively suppress the flow of heat into the rubble.

Heat from rain: Next, we consider the amount of melting due to heat introduced by rain during the summer after the winter snow cover has melted. In a maximum (and unrealistic) case, all the heat in the rain is transferred through the debris cover to the ice. From meteorologic data, summer rainfall in the Sierra Nevada is about 10% of total precipitation, and at 37° latitude near the crest, total annual precipitation is about 100 to 127 cm (40 to 50 in.; Rantz, 1969). Thus, average summer rainfall is about 10 to 13 cm. Over one summer, the total heat transferred by rain at 10°C is: (10 to 13) cm x 10°C cm^{-3} x 1 cal $°C^{-1}$ = 100 to 130 cal · cm^{-2}. This heat, if fully transferred to the ice, would melt somewhat less than a 2-cm thickness of ice at 0° C. This value is likely to be an overestimate because some rain will evaporate from the rubble before reaching the ice. For example, evaporation of only 2% (or 0.2 cm) of the rain would consume the entire amount of heat introduced by the precipitation. Such evaporation will tend to dissipate radiant heat accumulated at the surface of the debris during daytime heating. Furthermore, 10°C is probably too high for rain at ~3700 m altitude. Therefore, heat introduced by rain is negligible compared to other sources and will only be able to melt a relatively small amount of ice («2 cm) annually.

Heat from snow melt: Water from melting of overlying snow in the spring will introduce even less heat to the ice core. Snow typically melts from the surface downward as warm air and radiation introduce heat. The water thus produced will then have to pass through the underlying snow before reaching the glacier surface. This process will ensure that such snowmelt is at or very close to 0°C when it reaches the debris layer. Because glacier ice in the Sierra Nevada, both covered and uncovered, is temperate and therefore also at 0°C, the percolating water will introduce almost no heat to the underlying ice.

Other studies: Supporting the theoretical considerations I present above, several studies, both theoretical and empirical, have quantified the effects of debris cover on heating and ablation of glacier ice. Nakawo and Young (1982) and Bozhinskiy *et al.* (1986) identify and mathematically model the basic ablative processes acting on debris-covered glacier ice. Both studies show that linear increase in thickness of surficial debris greater than a few centimeters causes rates of ablation to decrease exponentially, such that sub-till melting practically ceases where surface debris is 2-3 m thick. Accordingly, only very minor ablation from percolating meltwater and rainwater should occur beneath such debris (Bozhinskiy *et al.*, 1986, p. 264).

Providing tests for these models, Østrem (1959) and Whalley (1974a) independently measured rates of ablation under small, thin plots of sand or gravel placed on bare-ice glaciers. Similarly, Loomis (1970) and Lundstrom *et al.* (1993) compared ablation rates to thickness of till cover in traverses, respectively, of a medial moraine on the Kaskawulsh Glacier, Yukon Territory, and the debris-covered terminal zone of Eliot Glacier on Mount Hood, Oregon. Their results, summarized in Figure 2.7, support the theoretical models, and demonstrate that 1) maximum rates of ablation, resulting from increased radiative absorption by the debris, occur with a debris thickness of about 1-2 cm; 2) rates of ablation under 2-4 cm of debris cover are approximately equal to those of bare ice; and 3) debris cover of 30 cm reduces ablation rates by about an order of magnitude. Østrem (1959, p. 229) and Potter (1972, p. 3025) estimate that for debris mantles of 1-2 m ablation is reduced by several orders of magnitude relative to that of

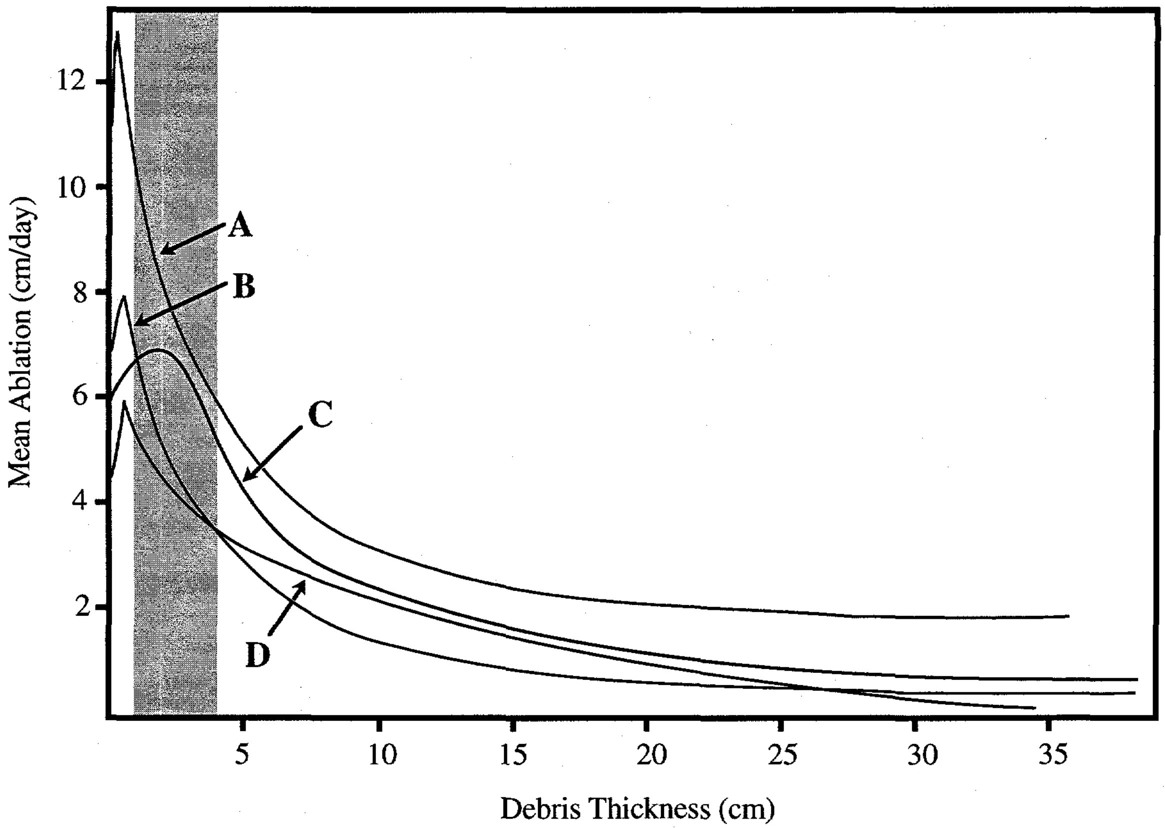


Figure 2.7: Graph summarizing observed rates of ice ablation as a function of thickness of debris cover. Shaded zone encompasses critical thicknesses of debris that decrease ablation relative to that of bare ice. Modified from Whalley (1974a); sources of curves: A, Whalley (1974a); B, Lundstrom et al. (1993); C, Loomis (1970); D, Østrem (1959).

bare ice, a conclusion supported by studies of the effects of a large, earthquake-induced debris slide on the Sherman Glacier in Alaska (Bull and Marangunic, 1968, p. 312).

Climatic and thermal conditions near Sierran rock glaciers are comparable to those of most of the studies listed above; therefore, their results can reasonably be applied to the rock glaciers I studied.

Implications of Debris-Covered Glaciers for the Glacial Record

The insulation provided by continuous debris cover on a glacier clearly allows an ablation area to extend to lower altitudes than those of equivalent bare-ice glaciers. Debris-covered glaciers and ice-cored moraines may thus persist well after climatic change thins or destroys exposed glacier ice. These conditions seem to be ubiquitous in Matthes deposits and probably affected glaciers of earlier advances.

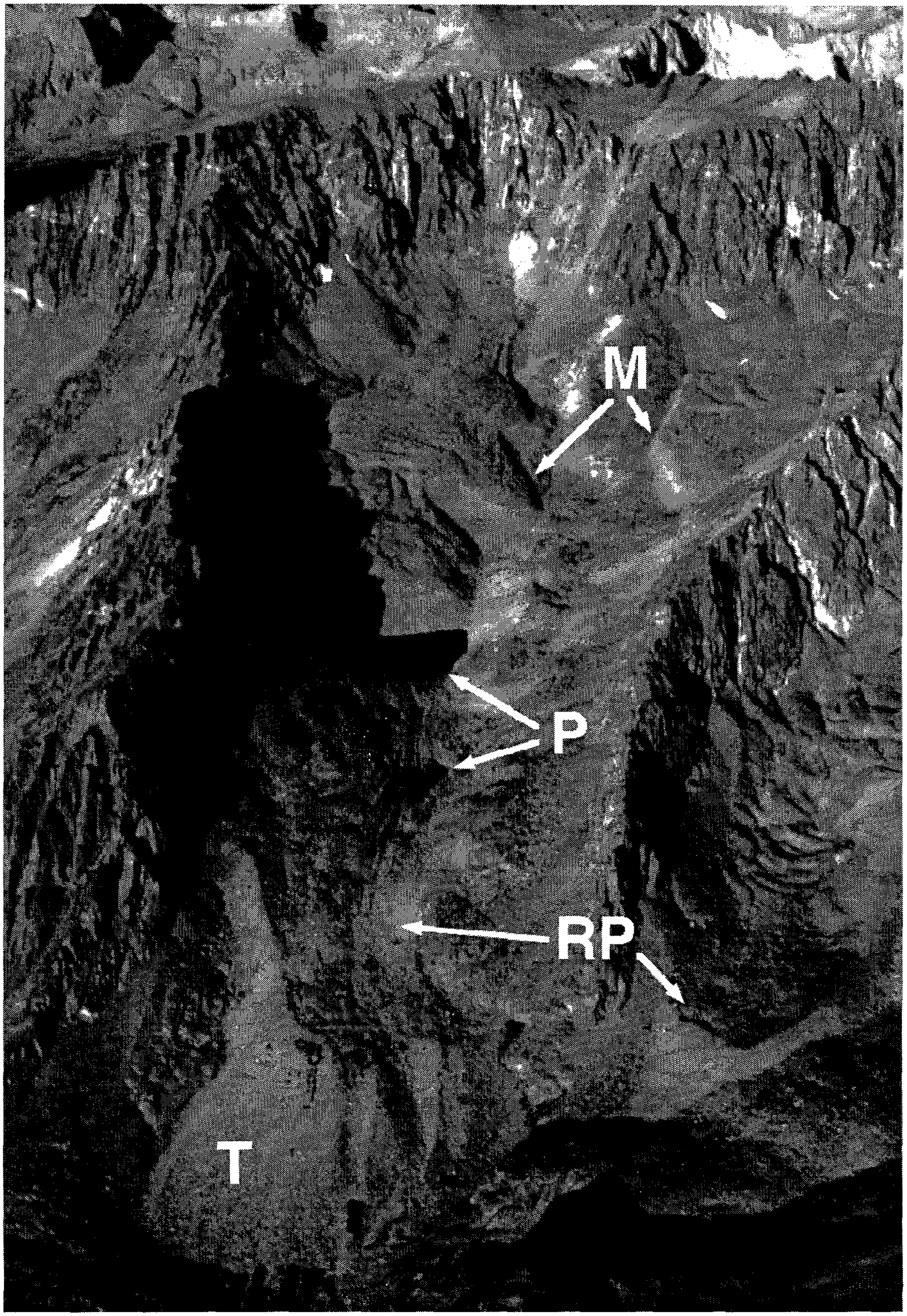
For example, reconstructed ELAs of Recess Peak cirque deposits in the type area (Birman, 1964) are about 150-200 m below modern ELAs (Burbank, 1991). By assuming a southward-rising ELA gradient along the Sierran crest for the Recess Peak advance, similar to that of late-Pleistocene and modern glaciers, Burbank found that ELAs estimated from deposits mapped as Recess Peak equivalents by Gillespie (1982) and Scuderi (1984) near the southern end of the range appear to be much lower than expected. Burbank (1991) argued that this apparent anomaly indicates that most of these moraines may have been deposited in the late Pleistocene before the Recess Peak advance (Yount *et al.*, 1982). I propose, instead, that many of these moraines in the southernmost Sierra Nevada, which occupy deep canyons and cirques where rockfall is common, may actually be Recess Peak debris-covered glaciers, while others possibly are former valley-wall rock glaciers. Thus, the large, apparently anomalous ELA depression may partly reflect an incorrect assumption that the Recess Peak glaciers lacked extensive debris mantles.

Because AARs and THARs used to estimate ELAs are derived empirically from modern bare-ice glaciers, the unusually large ablation areas of debris-covered glaciers would indicate significantly lower ELAs with these ratios than actually occur. My field observations of ~20 modern debris-covered glaciers indicate that AARs of such glaciers in the central and southern Sierra Nevada range from 0.2 to <0.1, and that THARs range from 0.6 to 0.8. These values differ significantly from those noted for extant bare-ice glaciers (AAR ≈ 0.65; THAR ≈ 0.4-0.5)(e.g., Meierding, 1982; Porter *et al.*, 1983).

Burbank (1991) applied a "bare-ice glacier" AAR of 0.65 to estimate Recess Peak ELAs along a crestal profile. However, the anomalously low Recess Peak deposits in the southern part of the profile display features typical of former debris-covered glaciers, including remnant longitudinal and transverse ridges, continuous dead-ice terrain up-canyon from terminal moraines, remnant thermokarst depressions, and an overall deflated appearance (e.g., Fig. 2.8). These deposits contrast markedly with most Recess Peak deposits elsewhere in the Sierra Nevada, which have small lateral and terminal moraines and minimal ablation drift, indicative of a brief glacier advance of relatively debris-free ice. Insulating debris cover on the southern Recess Peak glaciers would have allowed them to remain active longer and to reach considerably lower altitudes than did equivalent bare-ice glaciers elsewhere in the range. Thus, while Burbank's ELA estimates of the Recess Peak advance may be accurate for the bare-ice glaciers in the type area, the substantially lower values he obtained in the south may reflect debris-covered Recess Peak glaciers.

To test this hypothesis, I calculated ELAs for 40 Recess Peak glaciers along the same transect studied by Burbank (1991), using an AAR (= 0.15) more representative of debris-covered glaciers where appropriate. I used the same mapping and applied the same "filtering" criteria as did Burbank, but distinguished deposits of bare-ice glaciers from those of debris-covered glaciers on the basis of the landform characteristics noted

Figure 2.8: Oblique aerial photograph of remnants of Recess Peak debris-covered glaciers (RP) downstream from active Matthes debris-covered glaciers (M) in headwaters of Tinemaha Creek, eastern Sierra Nevada. Note relict morphology and thermokarst ponds (P), and generally deflated character typical of extinct debris-covered glaciers relative to modern equivalents. Terminal slope (T) of main Recess Peak debris-covered glacier stands at or less than angle of repose, bears some vegetation, and shows incipient gullying, all indicative of stagnation. View to west-southwest. August 24, 1992.


M
P
RP
T

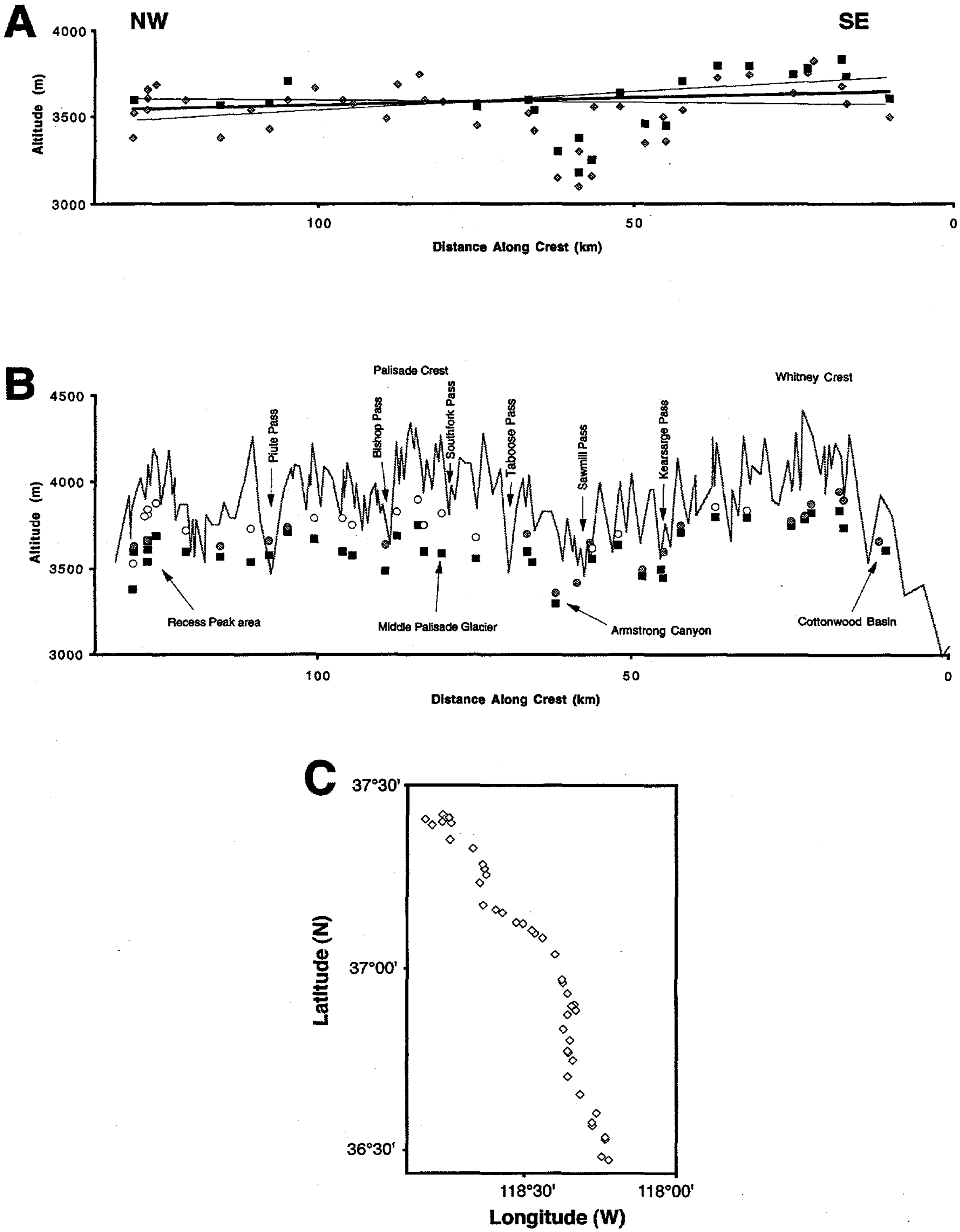
above. I also included ELA estimates of previously unmapped deposits located along the Sierran crest between the Recess Peak type area and Taboose Pass and between Independence Creek and Cottonwood Basin. All but two of the twenty-one Recess Peak glaciers south of Southfork Pass appear to have been mostly debris-covered (Fig. 2.9a), whereas a majority of those north of the pass had relatively little debris.

Recess Peak ELA Gradients in the Southern Sierra Nevada

The results demonstrate several interesting trends. First, the recalculations increase the ELAs of debris-covered Recess Peak glaciers an average of 112 m and shift several outlying values into closer accordance with those of adjacent bare-ice glaciers (Fig. 2.9a). However, regressing the corrected data still does not define a significant southward-rising trend (even within 90% confidence limits), as do ELA estimates for the late Pleistocene (Burbank, 1991; Gillespie, 1991), and substantial deviations from the regression remain.

Despite these deviations, Recess Peak ELAs vary relatively smoothly from the type area to the southern area without the discontinuity indicated in Burbank's (1991, Fig. 6) estimates. In comparing ELA gradients along the crest, Burbank argued that the Recess Peak ELA gradient should mimic the steady southward rise of the late-Pleistocene ELAs. In contrast, my data show that, along this transect, Recess Peak ELAs are largely independent of latitude; instead they generally appear to vary with altitude of the crest (Fig. 2.9b). In a gross sense, the ELA gradient is flat from the Recesses in the north to Southfork Pass, drops sharply near Sawmill Pass, climbs steadily to the south end of the Whitney Crest, then drops sharply into Cottonwood Basin. These variations occur at smaller scales also; south of Piute Pass, for example, ELAs decrease southward to a local minimum near Bishop Pass, climb abruptly along the Palisade Crest, then drop southward again, more-or-less paralleling the crest. Despite the "size-filtering" used by Burbank to

Figure 2.9. (A) NW-SE profile of Recess Peak ELAs along crest of Sierra Nevada from Recess Peak type area in north to Cottonwood Basin in south. Graph shows ELAs based on a "bare-ice" AAR of 0.65 (gray diamonds) and corrected values for debris-covered glaciers (AAR = 0.15; black squares). Corrections reduce variability about central trend and raise affected ELAs an average of 112 m. Although the corrected average gradient (bold line) rises slightly to southeast, the slope is not significant within 90% confidence limits (thin lines), and there are significant local departures from the trend. (B) Similar profile comparing best estimates of Recess Peak ELAs (black squares) with those of adjacent Matthes glaciers (circles). Gray circles indicate Matthes glaciers smaller than minimum-size criterion of Burbank (1991). Sierran crest shown by continuous line. Note that both Matthes and Recess Peak ELAs follow general altitude variations of crest. (C) Locations of Recess Peak and Matthes ELA data.



A
NW
SE
Altitude (m)
4000
3500
3000
100
50
0
Distance Along Crest (km)
B
Palisade Crest
Whitney Crest
Piute Pass
Bishop Pass
Southfork Pass
Taboose Pass
Sawmill Pass
Kearsarge Pass
4500
4000
3500
3000
Altitude (m)
Recess Peak area
Middle Palisade Glacier
Armstrong Canyon
Cottonwood Basin
100
50
0
Distance Along Crest (km)
C
37°30'
37°00'
36°30'
Latitude (N)
118°30'
118°00'
Longitude (W)

minimize the effects of local topography on ELAs, such influences apparently can overwhelm the regional climatic gradient for small (~1-3 km long) glaciers occupying deeply scoured cirques. Recalculating ELAs to account for the insulating effect of debris cover lessened but did not remove this apparent local orographic effect. Even assuming an AAR of 0.0 for the lowest glaciers does not eliminate it, because the cirque headwalls of these valleys are well below the regressed ELA gradient.

In the most extreme cases, a substantial local Recess Peak ELA depression remains in Armstrong and Sawmill canyons (Fig. 2.9a). The three deposits in Sawmill Canyon appear to be remnants of large valley-wall rock glacier complexes originating in taluses and thus may not represent former debris-covered glaciers, as I have defined them. Although apparently stagnant, as indicated by vegetated and sub-angle-of-repose fronts, the rock glaciers do not appear to have deflated substantially and therefore probably never had significant cores of glacier ice. As noted above, such rock glaciers can occur at substantially lower altitudes than can debris-covered glaciers. I have therefore removed these ELA values from Figure 2.9b, which retains my best estimates for Recess Peak ELAs. However, morphologic evidence indicates that the remaining outlier in the area, in Armstrong Canyon, is a former debris-covered glacier, and that the ELA depression there is real.

Assessment of Results

My best estimates for the Recess Peak ELAs along the profile provide a better fit to a regressed line than do the uncorrected values. The revised values cause average residuals to decrease from ~170 m to ~115 m, although r^2 only increases from 0.002 ($p < 0.78$) to 0.152 ($p < 0.02$). In addition, the slope of the line becomes significant within 95% confidence limits, rising 1.3 m km^{-1} to the south. The residuals progressively decrease if higher-order curves are fit to the data. With a 5th-order regression, the average residual is about 85 m, and the resulting curve strongly resembles a generalized topographic profile of the range crest. The resemblance probably reflects the dominance of topographic effects over regional climatic trends for cirque glaciers in the Sierra Nevada. The substantial rise in ELAs south of Kearsarge Pass may also partly reflect an increased rainshadow east of the Great Western Divide (Fig. 2.1).

To verify the trends for the Recess Peak ELAs, and to validate my mapping and reinterpretations of Recess Peak deposits, I estimated ELAs for Matthes-age glaciers and valley-floor rock glaciers in the same drainages. In the type area, Recess Peak ELAs typically are depressed approximately 160 m below the equivalent Matthes ELAs (Fig. 2.9b). If the deposits mapped as Recess Peak equivalents south of the type area, including those of Gillespie (1982) and Scuderi (1984), are actually remnants of an earlier, more extensive late-Pleistocene advance, as Burbank (1991) suggested, then there should be a distinct increase in separation of ELA values (ΔELA) between Matthes deposits and those mapped as Recess Peak. In fact, the results clearly show that Matthes ELAs vary consistently with those of Recess Peak glaciers, with even a slight convergence of values to the south (Fig. 2.9b). Regressing the Matthes ELAs against those of Recess Peak gives an r^2 value of 0.797 ($p < 0.001$), lending strong support to my mapping of Recess Peak deposits. The slight decrease in ΔELA to the south probably reflects the highly protected sites of these small Matthes glaciers, most of which do not meet Burbank's lower size limits. This observation further supports the notion that local topography becomes an increasingly important factor as the size of a glacier decreases. Thus, Burbank's (1991) "size-filtering" to weed-out glaciers not representative of regional climate is reasonable, but the lower size limit needs to be much larger (i.e., glaciers must terminate well beyond a cirque) to outweigh local topographic effects, at least in the Sierra Nevada.

The topographic dependence of ELAs of small glaciers occurring in deeply scoured terrain such as the southern Sierra Nevada indicates that accurate regional ELA gradients for such glaciers cannot be established unequivocally from just a few points. Such glaciers, typically up to 3-4 km long, owe their existence primarily to local orographic influences on snow accumulation (e.g., Matthes, 1942, p. 161). These factors, including cirque aspect, large amounts of avalanched and drifted snow, and shading below high cirque headwalls, act to increase snow accumulation substantially relative to that in adjacent, less sheltered locations. In the deep cirques and canyons of the southern Sierra Nevada, only the large Pleistocene glaciers apparently were unaffected by local orographic effects, because they extended well beyond the highly protected cirques near the crest.

The exceptionally low Recess Peak ELAs in and near Armstrong Canyon coincide with the eastward projection of the canyon of the South Fork Kings River. This nearly E-W-trending trough is exceptionally deep and wide, and may act as a storm conduit, allowing more precipitation to reach the crest of the range and thereby lowering ELAs (e.g., Porter, 1964, 1975a, p. 35, 1977; Burbank, 1991, p. 299). Burbank speculated on a similar deviation for late-Pleistocene glaciers near the canyons of the Kings River, although his data were too sparse to show it. Extending this idea northward to the San Joaquin River gorge, I would anticipate that Recess Peak ELAs there were lowered as much as or more than those near the Kings River (see Chapter 3). My data suggest that the effect was more pronounced during less extensive ice advances than during full-glacial conditions (also, Burbank, 1991; Gillespie, 1991), perhaps as a result of partial filling of the canyons by glacier ice during glacial maxima.

Large Late-Pleistocene Debris-Covered Glaciers in the Sierra Nevada

Reconnaissance mapping indicates that debris-covered glaciers were probably rare in the Sierra Nevada under full-glacial conditions (e.g., Tioga and Tahoe glaciations) and that the adjustments in ELA estimates that I used for Recess Peak debris-covered glaciers may be unnecessary for most large Pleistocene Sierran glaciers. Several factors probably combine to decrease the ratio of supraglacial debris to ice during maximum glaciations: covering of rockfall sources by more extensive snow and glacier ice, reduced rockfall because of fewer freeze-thaw cycles and decreased chemical weathering in the highest parts of the range, and increased rates of snow accumulation and glacier flow. Thus, continuous debris cover might occur only near the glacier termini where flow rates are slow and ablation is highest. Transitory exceptions may have occurred, such as massive, earthquake-induced rockfalls that can bury much of a glacier's ablation area (i.e., several Alaskan glaciers after the 1964 Anchorage quake; Post, 1967). However, evidence for such events might be difficult to recognize.

In the Rocky Mountains, Lundstrom (1991) cites the occurrence of massive and extensive late-Pleistocene lateral moraines as evidence that debris-covered glaciers were common there during full-glacial times as a result of increased rockfall. He concluded that the thermal shielding provided by the debris cover allowed the glaciers to grow larger than if they had been clean, and he implied that late-Pleistocene ELA estimates for the region (using AAR ≥ 0.6) were 30-50% too low. Using the morphologic criteria established for the Recess Peak stade, evidence is sparse for large, late-Pleistocene debris-covered glaciers in the Sierra Nevada (e.g., bare bedrock dominates most canyons between the Recess Peak and Tioga end/recessional moraines). The large lateral moraines cited by Lundstrom as evidence for high englacial debris loads can also result from long duration of construction and repeated addition of debris during successive glaciations. The two most likely sites of Pleistocene debris-covered glaciers in the southeastern Sierra Nevada are Armstrong and Division canyons. Other likely sites of late-Pleistocene debris-covered glaciers farther north include Sherwin Creek and a small drainage on the eastern flank of Mt. Wood (Fig. 2.1) In none of these locations, however, are the estimated Pleistocene ELAs substantially different from those of nearby drainages (Gillespie, 1991). Therefore, although these glaciers transported much debris, it apparently was either not thick enough or sufficiently continuous to insulate the surface effectively, or ice accumulation was significantly lower than in nearby drainages.

CONCLUSIONS TO CHAPTER 2

1. Detailed and reconnaissance investigations indicate that many Matthes (Little Ice Age) moraines and most active valley-floor rock glaciers in the southern Sierra Nevada consist of glacier ice covered by continuous but relatively thin debris, and that such rock glaciers are more accurately described as debris-covered glaciers. This finding contradicts the view that *all* rock glaciers are permafrost phenomena (e.g., Haeberli, 1985; Barsch, 1988).

2. Theoretical and empirical considerations of the effects of such debris cover on cirque glaciers demonstrate that insulation provided by ~1 m of debris will decrease rates of ablation by at least an order of magnitude. This decreased ablation allows debris-covered glaciers to reach substantially larger sizes and lower altitudes than nearby bare-ice glaciers, and to be less sensitive to climatic warming.

3. Accumulation-area ratios and toe-to-headwall-altitude ratios used in glacial, climatologic, and geologic studies to estimate ELAs of extinct glaciers, which are derived empirically from modern bare-ice glaciers, may yield incorrect results if the deposits of debris-covered glaciers are not distinguished. My study of anomalously low pre-Matthes Recess Peak deposits in the southern Sierra Nevada indicates that many of these deposits are actually remnants of debris-covered glaciers. Our criteria for distinguishing such

deposits from remnants of bare-ice glaciers include extensive ablation drift upcanyon from the terminal moraine, residual longitudinal and transverse ridges in the drift; residual thermokarst features, and a general concave, deflated character. Evaluating ELAs for these glaciers using an AAR (or THAR) more appropriate for such glaciers (AAR $\approx$ 0.15; THAR $\approx$ 0.7) decreases the deviations from the general southward-rising ELA trend and explains some of the variance.

4. Comparison of adjoining Recess Peak and Matthes ELAs to adjacent crestal altitudes along a 120-km reach of the Sierran crest indicates that small glaciers (up to 3-4 km long) have formed in deeply cut cirques mainly because of greater snow accumulation due to local orographic influences, and do not directly reflect regional climatic gradients. ELAs estimated for such glaciers and former glaciers, especially from statistically or geographically limited samples, may be unreliable.

5. Debris-covered glaciers were probably relatively rare in the Sierra Nevada under full-glacial conditions because of increased rates of snow accumulation and decreased rates of supraglacial rockfall.

ADDENDUM TO CHAPTER 2

Publication of this chapter in *Quaternary Research*, with Malcolm Clark and Alan Gillespie as co-authors (Clark *et al.*, 1994a), precipitated a rather heated commentary from M. Jakob of the Department of Geography, University of British Columbia (Jakob, 1994; Appendix 1). Jakob, who was a student of the prime advocate of the permafrost-only concept of rock glacier origin (W. Haeberli), strongly disagreed with our findings on Sierran rock glaciers. In this section, I present a modified version of our reply (Clark *et al.*, 1994b) to Jakob's criticisms of the paper.

I include this section because it addresses succinctly many of the major problems with the arguments presented by those who do not accept the possibility of a glacial origin for at least some rock glaciers. For much of the past decade, these "permafrost-only" advocates have dominated the scientific discussion of rock glaciers and have generally dismissed the glaciogenic model out of hand. Several recent studies, including my own, have reintroduced the glacier ice-cored rock glacier as a valid concept, and it seems appropriate here to address the contrarian arguments that have been raised by Jakob concerning my work.

M. Jakob's comment (Jakob, 1994; Appendix A) reflects the view that rock glaciers are exclusively a periglacial phenomenon. This belief is based mainly on inferences from indirect (geophysical) data. My work, in contrast, is supported by direct field observations of Sierran rock glaciers for which the clearest and simplest explanation is a glacial rather than periglacial (permafrost) origin. Field data indicate that the rock glaciers I observed in the Sierra Nevada are cored by glacier ice, that other similar landforms in the range are probably cored by glacier ice, and that by analogy many other rock glaciers elsewhere in the world may also be glaciogenic. Despite Jakob's inference, Clark *et al.* (1994a) do not insist that *all* rock glaciers have a glacial origin, and we even suggest that periglacial and glacial "rock glaciers" may coexist in the same valley. I therefore disagree with Jakob's claim that we "oppose" the view of rock glaciers as periglacial phenomena. Rather, I dispute Jakob's insistence that all rock glaciers are wholly periglacial. This distinction is important; in my view, rock glaciers are a morphologic form, not a singular genetic form.

Jakob further implies that Clark *et al.*'s (1994a) conclusion that some rock glaciers are glaciogenic is isolated amongst a well supported consensus for the periglacial model. This is not the case: direct observations from many thorough studies other than mine indicate continuity of glacial ice beneath the surface debris of some rock glaciers (e.g., Lliboutry, 1953, 1990; Outcalt and Benedict, 1965; Potter, 1972; White, 1971, 1976; Gardner, 1978; Ellis and Calkin, 1979; Jackson and MacDonald, 1980; Benedict *et al.*, 1986; Martin and Whalley, 1987; Calkin *et al.*, 1987; Johnson and Lacasse, 1988; Whalley, 1992; Whalley and Martin, 1992).

Jakob's main criticisms are that (1) observations in Clark *et al.* (1994a) did not demonstrate a continuous core of glacier ice beneath the surface debris, and (2) the absence of "absolute" dates on the deposits that I correlate to the Recess Peak advance invalidates my conclusions concerning regional snowline reconstructions for that period. I disagree with Jakob and find that his arguments are based on misinterpretations of both Clark *et al.* and others' papers, rely on undocumented assertions, or favor arbitrary terminology in lieu of process.

Continuity of Glacier Ice

Within Jakob's first criticism are imbedded two questions: (a) is there a continuous ice core in the Southfork Pass rock glacier? and (b) is the ice glacial? Trenching and seismic surveys, which might otherwise document subsurface ice (e.g., Potter, 1972), are difficult or impossible in the coarse, loose surface rubble typical of Sierran rock glaciers. A continuous ice core within the Southfork Pass rock glacier is clear, however, from the presence of the ice-walled melt pond and the absence of morphologic discontinuities in the surface debris downstream of the pond. Strong variations in the rubble thickness or rubble content of the underlying core (i.e., the clear ice exposed in the melt pond is not representative of rock glacier as a whole) would change the rheology of the rock glacier and would be manifest as (1) textural variability of surface morphology (e.g., Loewenherz *et al.*, 1989), and (2) velocity irregularities. As noted in Clark *et al.* (1994a), there is no significant textural variability other than that related to changes in slope. Similarly, surface velocities in a transect up the center of the rock glacier (Fig. 2.4) also primarily reflect changes in slope. Furthermore, velocities increase dramatically at the upper end of the rock glacier as would be predicted if the rock glacier is actually the ablation zone of the headwall glacier, as Clark *et al.* (1994a) conclude.

If the rock glacier were till remobilized from a moraine by addition of permafrost ice, as envisioned by the periglacial model (e.g., Haeberli, 1985, Fig. 4; Barsch, 1988, Fig. 4.9) and required by Jakob, then maximum velocities should occur near the center of the debris-covered area where the mass flux would be greatest (Haeberli, 1985, p. 88-92). This is clearly not the case at Southfork Pass. Therefore, the ice core of the Southfork Pass rock glacier is most likely continuous. A continuous ice core within the nearby Goethe Glacier moraine and rock-glacier complex (Fig. 2.5b) was also documented by M. Loughman (C. Wahrhaftig, written communication, 1993), who discovered a longitudinal tunnel in clear, banded ice that traversed the length of the complex.

As to the origin of the ice, Jakob takes issue with the conclusion that the layered ice exposed in the wall of the melt-pond in the Southfork Pass rock glacier is glacial and implies that it formed as subtill segregation ice. However, as he admits, such ice forms sub-horizontally (or parallel to the surface) and has complex, discontinuous layering (e.g., Pollard and Dallimore, 1988; Mackay and Dallimore, 1992, Fig. 4). I reemphasize that the layered ice in the Southfork Pass rock glacier dips uniformly up-glacier at ~40°, that the layers are continuous and unconformable with the surface, and that most of the layering comprises alternating milky and transparent bands of clear ice, with no intervening silt layers. Such layering is easily explained by glacial accumulation and rotation due to down-valley flow, but is much more difficult or impossible to explain using the periglacial model. Furthermore, one can clearly observe the ice of the accumulation zone only 100 m upslope from the pond being progressively buried by rockfall from the cirque headwall; the ice continues under the debris cover, and does not overlie it. Because these field relationships are obvious, I felt it unnecessary to include petrographic studies of the ice, as suggested by Jakob. His further assertion that, even if the ice is glacial, it is not representative of the rock glacier is not supported by my observations. The combined rheologic, glaciologic, and geomorphic evidence is most simply explained by a continuous glacial core within the Southfork Pass rock glacier. These findings are in accord with those of the many studies mentioned earlier.

Jakob asserts that the model of exclusive periglacial origin, which contradicts our findings, is confirmed by geophysical or geochemical measurements on rock glaciers around the world, and he criticizes us for not making such measurements. These measurements, however, are indirect observations, none of which distinguishes glacier ice from periglacial ice in a fundamental way (e.g., Whalley and Martin, 1992; Whalley and Azizi, 1994). Considering these ambiguities, the difficult access to rock glaciers in the Sierra Nevada, and the clarity of our direct observations (Clark *et al.*, 1994a), my co-authors and I therefore did not find it useful to duplicate such measurements in our study.

Jakob's arguments concerning the presence or absence of permafrost in the Sierra Nevada seem misguided. Of primary importance, the presence of permafrost in a region does not preclude rock glaciers of glacial origin, because glaciers and debris-covered glaciers can certainly exist in areas with permafrost. In any case, the claim that rock glaciers may be the only morphologic expression of alpine permafrost is circular and depends on the belief that all rock glaciers have a periglacial origin. Jakob's contention that debris-covered glaciers do not display transverse ridges and other "typical rock-glacier" morphologies is also circular, since he excludes the possibility that at least some rock glaciers are debris-covered glaciers (e.g., Whalley and Azizi, 1994, p. 49). Jakob's contention that the near-freezing temperature of the saturated till in the terminus of the Southfork Pass rock glacier is evidence for permafrost is unwarranted, because meltwater from a temperate glacier-ice core will also be at the freezing point. Likewise, the process of Balch ventilation is not restricted to permafrost regions, and cannot be used as an argument for the presence of such.

Other contentions by Jakob regarding our findings on rock glaciers seem to reflect semantics rather than process. By definition, permafrost is any ground frozen for more than a year; glaciers, which technically meet this definition, are generally, and arbitrarily, excluded (e.g., Washburn, 1980, p. 21; Barsch, 1977, p. 232). Thus, a rock glacier formed primarily by avalanching snow and rock (e.g., some valley-wall rock glaciers) may be, *sensu stricto*, periglacial (i.e., not glacial) in origin, but the same depositional processes also feed true glaciers in adjacent cirques. Also, superimposed ice can be added locally to a glacier-ice core by refreezing meltwater to the surface of a glacier or rock glacier (e.g., Sugden and John, 1976, p. 16; Whalley, 1983, p. 1399). In such transitional rock glaciers the distinction with glacier formation is blurred, and intermediate forms between periglacial rock glaciers and debris-covered glaciers certainly exist. Similarly, Jakob's contention with our use of "thermokarst" seems misdirected. Although the term strictly refers to melting of permafrost ice, from a process viewpoint, it is equally applicable to melting debris-covered glaciers, and has commonly been used in this manner (e.g., Barsch, 1987, p. 42; Barsch and King, 1989, p. 153; Kirkbride, 1993, p. 234).

Age and Significance of Sierran Rock Glaciers

There are two problems with Jakob's estimates of the age of the Southfork Pass rock glacier. First, I correlate it to the Matthes (Little Ice Age) advance, not the Recess Peak advance as he mistakenly states. Second, he extrapolates linearly a motion measurement near the terminus (Clark *et al.*, 1994a) to calculate an age of 8000 to 16,000 yr for the rock glacier, probably an order of magnitude too great. This age estimate disregards substantial changes in climate that have occurred in the Sierra Nevada during the Little Ice Age. In particular, receding glaciers and stranded lateral moraines demonstrate that rates of snow accumulation, and therefore glacier flow rates, were much higher in cirques during most of the period, and have decreased substantially during the past 100 years. Therefore, modern glacier flow rates may seriously underestimate those typical of the past 1000 years. Jakob's Pleistocene age for the rock glacier requires the Little Ice Age glacier in the cirque headwall to have somehow underthrust the older and denser rock glacier, a process for which there is no evidence.

Jakob further contends that Clark *et al.* (1994a) "assume" that pre-Matthes cirque deposits in the southern Sierra Nevada are Recess Peak in age. In fact, as our paper states, we correlated the deposits, and their Matthes equivalents, to the type areas using well established methods common to most glacial-geologic studies, including those of Burbank (1991) and Yount *et al.* (1982). Furthermore, we made no conclusions as to the actual age of the Recess Peak event, and therefore contradicted neither Yount *et al.* (1982) nor Burbank (1991) on the subject. Numeric ("absolute") dates are thus immaterial to our findings. I have discussed at length the results of our study with J. Yount, R. Burke, P. Birkeland, and D. Burbank, and none disagrees with the basic conclusions of the paper.

The observations my co-authors and I presented in the original paper (Clark *et al.*, 1994a) indicate a glacial origin for modern valley-floor rock glaciers in the Sierra Nevada, and that deposits of similar but earlier debris-covered glaciers in the southern part of the range correlate well with Recess Peak moraines in the type area to the north. We do not presume a glacial origin for all rock glaciers, as Jakob implies. The conclusions in our paper remain valid, particularly if one considers the operant processes in lieu of arbitrary nomenclature, and avoids unfounded or irrelevant concepts about the relation between permafrost and alpine glaciation.

CHAPTER 3: LATE-GLACIAL AND HOLOCENE CIRQUE GLACIATION IN THE SIERRA NEVADA

INTRODUCTION TO CHAPTER 3

Despite more than a century of research on major glaciation in the Sierra Nevada, few studies discuss the less-extensive glacial deposits that postdate the Tioga advance. The earliest workers in the range (e.g., LeConte, 1873; Muir, 1873; Russell, 1885; Gilbert, 1904) described moraines being formed by the modern glaciers, and Russell (1885, p. 322) first noted a small, "ancient" moraine a short distance below a modern glacier near Mono Lake. Matthes (1940, p. 399; 1960, Fig. 39) observed that two sets of small moraines typically occur immediately downstream from many modern glaciers in the Sierra Nevada, and considered them both to be equivalents of the historical advances in the European Alps (Matthes, 1940, 1941). Yet these apparently young deposits were largely ignored in geological reports until Birman (1964) published a detailed transect of glacial deposits across the central part of the range. In his study, Birman distinguished three post-Tioga glacier advances, all of which he felt were late-Holocene age: Hilgard, Recess Peak, and Matthes (Little Ice Age), in decreasing age.

In this chapter, I introduce new data on the age and climatic significance of the Recess Peak advance and its relation to earlier and subsequent advances. In particular, I reconstruct 64 Recess Peak and adjacent Matthes glaciers and their equilibrium-line altitudes (ELAs) along a 400-km profile of the Sierran crest, from the southernmost deposit at Olancha Peak to the northernmost well-documented deposit, near Lake Tahoe (Fig. 3.1). I also present a series of high-resolution, minimum-limiting radiocarbon dates from lake sediments next to Recess Peak moraines in four widely separated locations. My findings demonstrate that Recess Peak deposits represent a small but significant Late-Pleistocene climatic event that led to glacier readvance in the Sierra Nevada, possibly

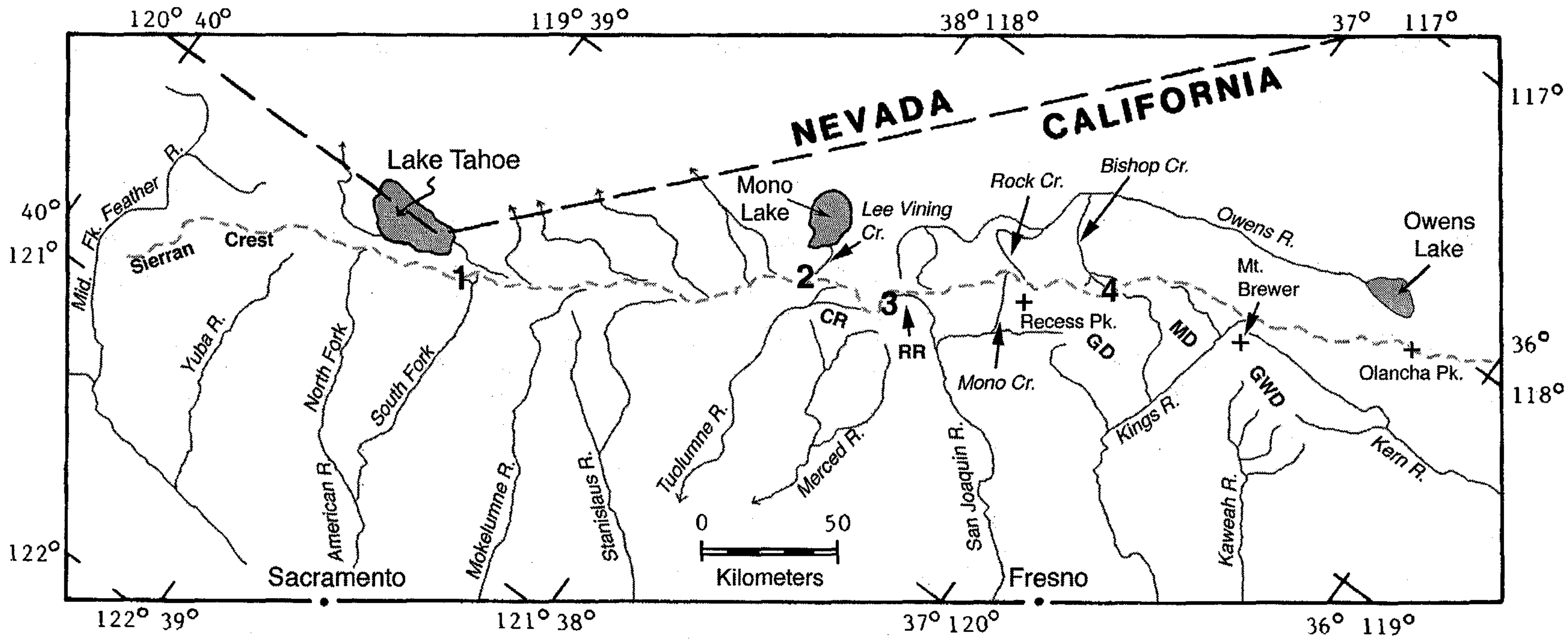


Figure 3.1: Map of Sierra Nevada, California. Recess Peak and Matthes deposits occur along the range crest from southern Lake Tahoe to Olancha Peak. Thin black lines show major drainages and rivers discussed in text; dashed gray line shows crest of Sierra Nevada. Numbers refer to four coring sites discussed in text: 1 - tarn above Lake Aloha, South Fork American River; 2 - Green Treble Lake, Lee Vining Creek; 3 - upper Garnet Lakes, Middle Fork San Joaquin River; 4 - Baboon Lakes, Middle Fork Bishop Creek. CR, Cathedral Range; GD, Goddard Divide; GWD, Great Western Divide; MD, Monarch Divide; RR, Ritter Range. Modified from Wahrhaftig and Birman (1965).

related to Tioga retreat, and that this readvance predates the North Atlantic Younger Dryas event. Furthermore, these results require that glaciers were either quite small or absent in the Sierra Nevada during the Younger Dryas period (~12,800-11,500 sidereal yr B.P.; Johnsen *et al.*, 1992; Alley *et al.*, 1993; ~11,000-10,000 ^{14}C yr B.P., Mangerud *et al.*, 1974), and indeed, have not reached beyond the Matthes ice limits during the entire ~12,000-yr period that appears to separate the two advances. Other work (Birkeland *et al.*, 1976; Yount *et al.*, 1982; Clark and Clark, 1995) indicates that moraines of Birman's (1964) "Hilgard glaciation" are probably related to Tioga recession and do not represent a distinct advance.

Previous Studies

The Recess Peak and Matthes advances were originally defined on the basis of relative weathering (RW) parameters and topographic position of their respective moraines (Birman, 1964). The type deposits are near the headwaters of Mono Creek (Fig. 3.1). As described by Birman, Recess Peak moraines typically look "fresh", have ~70-90% unweathered boulders with substantial lichen cover, and have slightly developed soils. Moraine slopes are stable, with only minor incision and post-depositional erosion (Figs. 3.2, 3.3; Birman, 1964). Scattered small trees and shrubs occur on the moraines, but vegetation is generally scarce (Fig. 3.3). As Birman (1964, p. 54) noted, these features are very different from those of nearby Matthes deposits, which are typically still unstable, unweathered, have no significant gullying or vegetation cover, and possess little or no lichen cover. Many Matthes moraines are still associated with thinning or retreating modern glaciers. The distinction between the two deposits is useful because the Recess Peak deposits invariably lie close to, and are locally partially buried by, the Matthes moraines. Both are generally within 1-3 km of the cirque headwalls (e.g., Fig. 3.2). The outer set of “historic” moraines described by Matthes (1940) almost

Figure 3.2: Oblique aerial photo of northeast side of Mt. Brewer, southern Sierra Nevada (Fig. 3.1), showing typical relationship of the two major cirque deposits in the range: Matthes (Little Ice Age) moraine under cirque headwall, and Recess Peak moraine below it in lower foreground. A small, shaded, probably stagnant modern remnant of Matthes glacier occupies highest part of cirque. Note stable but uneroded appearance, small volume, and closely spaced multiple crests of Recess Peak moraine, in contrast with unstable, more voluminous, and simple structure of Matthes moraine. Extensive bare granitic bedrock and absence of any moraines between Recess Peak and Matthes moraines shown here is typical throughout the Sierra Nevada. Photo by M. M. Clark, September 23, 1977.

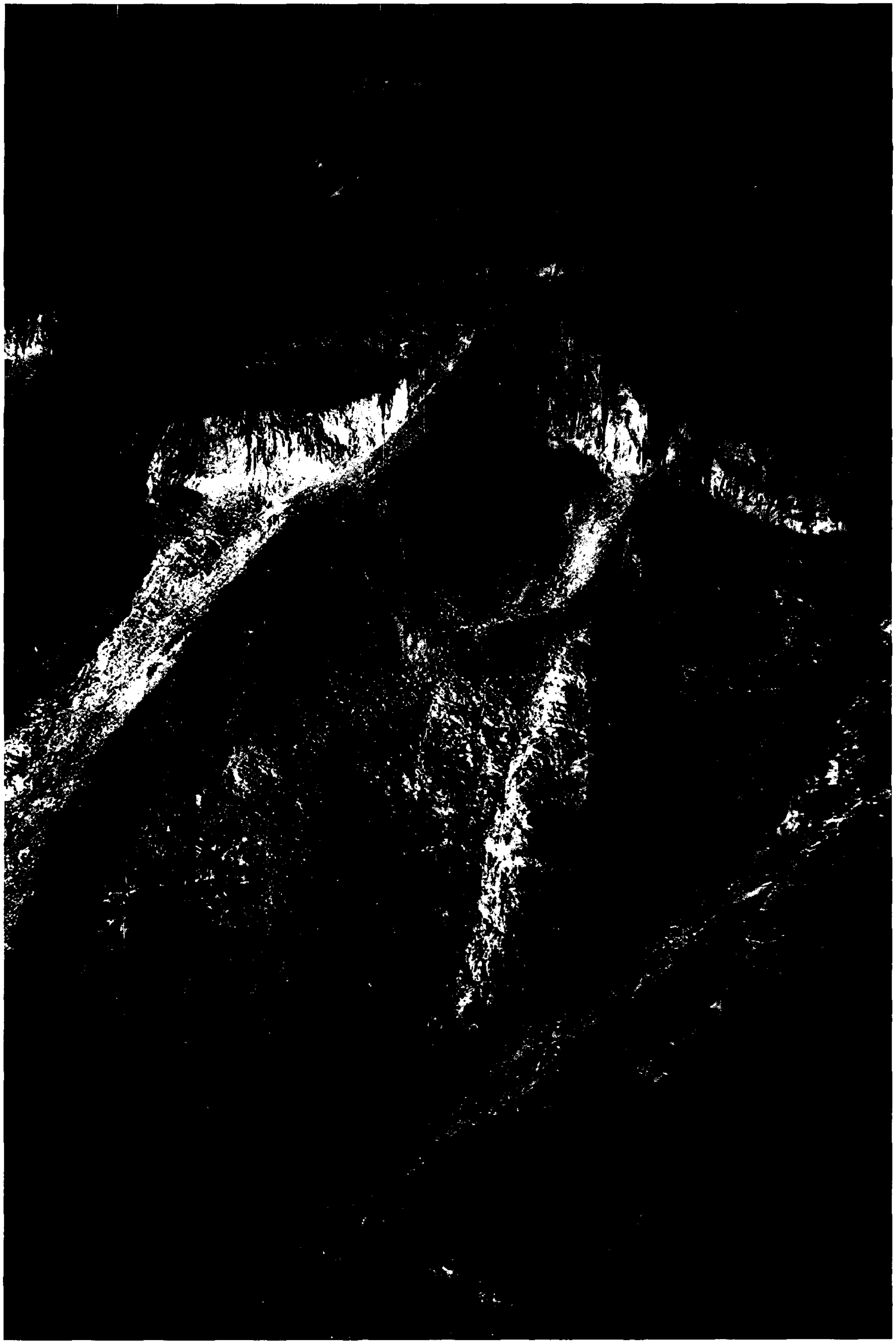


Figure 3.3: View to SSE (upvalley) of left-lateral Recess Peak moraine above Baboon Lakes (see Fig. 3.10). Former glacier occupied left side of photo, flowing from peaks visible in background. This moraine is typical of most Recess Peak moraines, with low volume, fresh but stable boulders and slopes, incipient soil formation, and small scattered clumps of trees or shrubs (Whitebark pines here).

certainly correspond to the Recess Peak advance. Small moraines mentioned by Russell (1885, p. 322) and Matthes (1960, Fig. 39) also correlate with the Recess Peak deposits.

Tephra stratigraphy provides the best numerical age limit for the Matthes advance. Wood (1977, p. 91) noted that his "tephra 1," a Mono Craters tephra erupted 720 ± 60 ^{14}C yr B.P. (650-680 cal yr B.P.), overlies all glacial deposits in the Ritter Range (Fig. 3.1), as mapped by Janda (1966), except the Matthes moraines. Yount *et al.* (1982) observed that what is probably the same tephra (with a minimum limiting radiocarbon date of 700 ± 60 ^{14}C yr B.P. (570-670 cal yr B.P.) covers N3 (Recess Peak) deposits but does not blanket N1 (Matthes) moraines in the Recess Peak area. Thus, as Wood (1977) concluded, Matthes glaciers in the central Sierra Nevada reached maximum positions after ~700 ^{14}C yr B.P. (~650 cal yr B.P.).

The age of the Recess Peak event has been disputed (Fig. 3.4). Birman (1964) originally proposed that RW criteria and correlation with the glacial records of the Rocky Mountains and European Alps indicated that the Recess Peak advance occurred during the last millennium, most likely during the 16th or 17th century AD. On the basis of lichenometry, Curry (1969, 1971) subsequently argued that Recess Peak moraines formed during two mid-Neoglacial advances occurring about 2700 and 2000 years ago. Using a lichen growth curve based on Curry's (1969) work, Scuderi (1987b) proposed a similar sequence. However, both Curry's (1969, p.19) and Scuderi's (1987b, p. 226) curves were not well controlled beyond historic times (~100 yr); in fact, Curry's long-term growth rates are based on lichens on moraines that he assumed were ~1000, ~2000, and ~2600 yr old, introducing a circularity in his (and Scuderi's) age assignments, in that all the control moraines probably correlate with the type-Recess Peak deposits.

Yount *et al.* (1979, 1982) compiled multiple RW measurements on Recess Peak and Matthes deposits in the type region (Fig. 3.1). On the basis of these measurements, tephrochronology, and several mid- to late-Holocene radiocarbon dates from soil pits

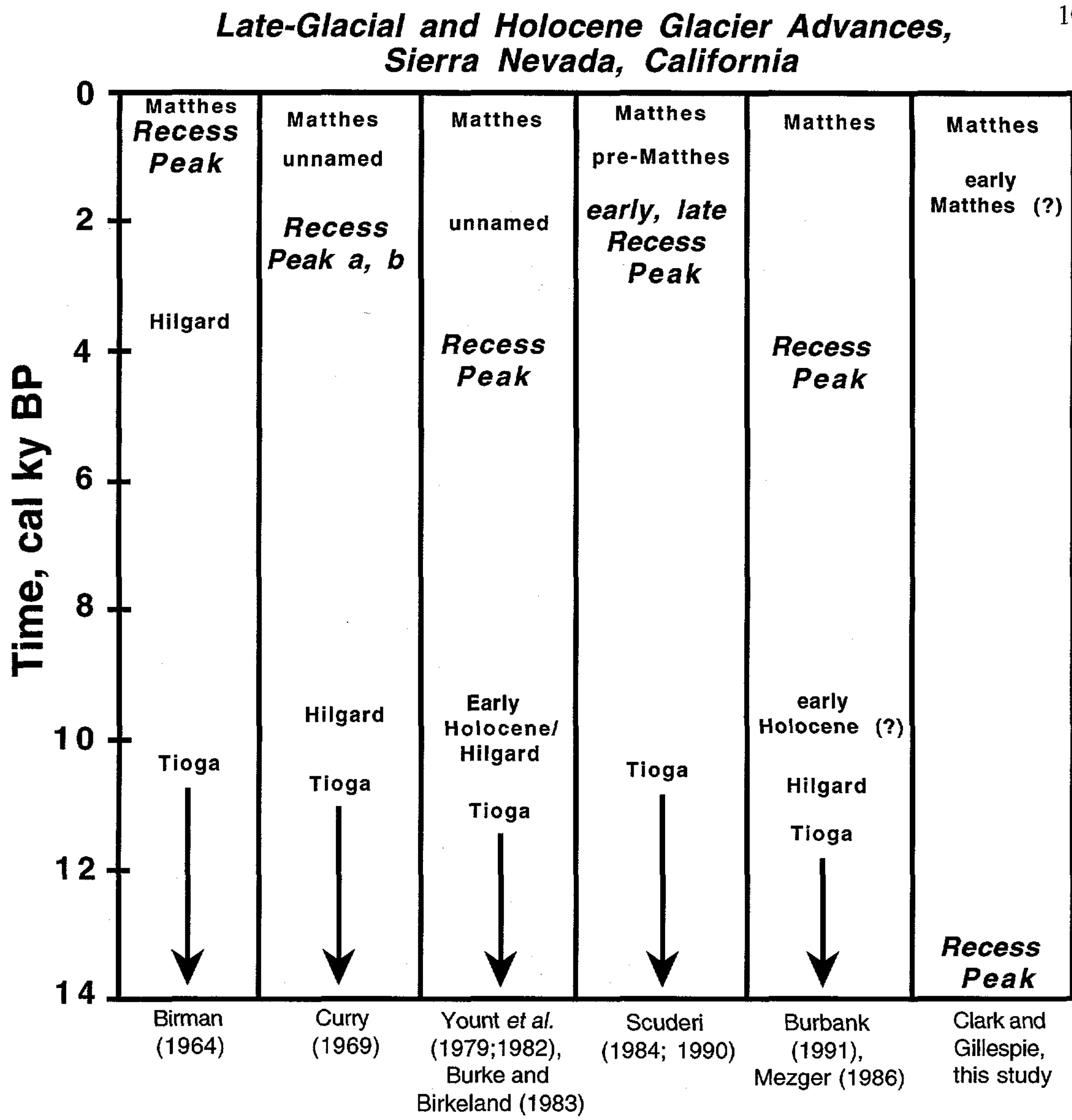


Figure 3.4. Summary of published latest-Pleistocene and Holocene stratigraphic names and ages assigned to moraines in the Sierra Nevada. Recess Peak advance emphasized. "Early Matthes (?)" in rightmost column signifies possible initiation of Matthes or pre-Matthes Neoglaciation prior to Matthes maximum. Supposition based on rate of flow of Matthes rock glaciers, and relative weathering observations of Yount *et al.* (1982).

within a set of Recess Peak moraines, they concluded that at least some of the deposits mapped as Recess Peak by Birman (1964) were probably late Pleistocene or early Holocene in age. However, the temporal and spatial relationships between possible late-Pleistocene and Neoglacial (i.e., post-hysithermal) deposits remained unclear and Yount *et al.* (1982) retained "Recess Peak" as an early Neoglacial advance (Fig. 3.4). Deposits possibly correlative with Recess Peak moraines have been mapped locally elsewhere in the range (e.g., Birkeland, 1964; Gillespie, 1982; Scuderi, 1984, 1987b; Mezger, 1986; Mezger and Burbank, 1986; Whiting, 1985, 1986), but the regional implications of the deposits were generally not considered. Burbank (1991) pointed out that reconstructed ELAs based on Recess Peak deposits mapped in the southern end of the range appeared discordant with those in the type area. He surmised that the deposits actually represent two separate populations, one representing a late-Pleistocene advance in the south and the other a Neoglacial advance in the north. Subsequent work by Clark *et al.* (1994a, b) demonstrates that most of the deposits are from a single population, and that the discordant ELAs reflect strong local influences of debris cover and orographic factors on the formation and maintenance of small cirque glaciers such as those of the Recess Peak advance. The results I present in this chapter build on initial work by Clark *et al.* (1994a; Chapter 2, this dissertation) in defining the climatic regime that formed the Recess Peak glaciers, and provide new information on the timing and extent of late-glacial and Holocene glacial advances in the Sierra Nevada.

A separate "glaciation," the Hilgard, defined by Birman (1964) as a distinct advance occurring between Tioga and Recess Peak advances, has been extensively referenced in glacial studies as a late-glacial or Holocene event. Birman distinguished these deposits primarily on the basis of relative weathering estimates. Clark (1976) and Clark and Clark (1995), however, reevaluated Hilgard deposits in the type area and elsewhere in the range and failed to find these distinctions. Their evidence instead indicates that type Hilgard moraines are probably recessional moraines of the Tioga stade, as previously suggested by Birkeland *et al.* (1976, p. 289), and that this term should be abandoned. I briefly discuss the implications of these findings later in this paper.

METHODS

This study has two main components: (1) estimates of ELA depression, and thereby relative climate, associated with the Recess Peak and Matthes glacier advances, and, (2) dating of sediments from alpine lakes associated with Recess Peak moraines in order to gain new limits on the timing of cirque glaciation in the range. The first component involved mapping and reconstructing the extent of glacier ice during the two advances, and the second involved coring lakes in which sedimentation began after the Recess Peak glaciers disappeared. All mapping and correlations I use in this study are my own, including those from areas previously mapped by others (Birman, 1964; Janda, 1966; Yount *et al.*, 1982; Scuderi, 1984; Whiting, 1985; Mezger, 1986; Gillespie, 1982, 1991; M. Clark, unpublished data, 1993). However, my mapping does not generally differ greatly from that of the earlier studies.

Mapping

Mapping methods for this study followed those described in Chapter 2 and in Clark *et al.* (1994a, p. 140). I mapped Recess Peak and Matthes deposits in 64 cirques near the Sierran crest from 36°16’ N, 118°07’ W to 38°52’ N, 120°10’ W. This interval spans the documented N-S extent of Recess Peak and Matthes deposits in the range, although some small protalus ramparts and cirque deposits mapped near the north end of Lake Tahoe (Birkeland, 1964, p. 821) and near the divide between the Feather and Yuba rivers (D. Clark and M. Clark, unpublished data) may also correlate with type Recess Peak deposits. Reconnaissance mapping indicates that small moraines formed in the northeastern cirque of Patterson Mountain in the Warner Mountains, northeastern California, may represent more-northerly expressions of the Matthes and Recess Peak advances (D. Clark, *unpublished data*); other northern equivalents outside California may also exist (e.g., Waitt *et al.*, 1982).

I combined field investigations and measurements with photo-interpretation of 1:16,000-scale color and 1:60,000 and 1:80,000 black-and-white aerial photographs to reconstruct the extent of the former glaciers. I distinguished moraines and other deposits of Matthes and Recess Peak advances on the basis of stratigraphic position, morphologic character, lichen cover, and general weathering characteristics. Matthes deposits are unweathered and commonly unstable, exhibit no significant erosion, soil development, or vegetation cover, and generally occur next to modern glaciers or snowfields. Matthes moraines commonly appear to be very bulky, generally reflecting the presence of a glacier-ice core beneath the morainal debris; in many, especially in the southern part of the range, the ice cores continue to flow, forming active rock glaciers (or debris-covered glaciers) with oversteepened front slopes (Clark *et al.*, 1994a, Fig. 5). These youthful features allow Matthes deposits to be readily identified. In contrast to the Matthes moraines, Recess Peak moraines typically have low volume, are generally stable, show incipient gullying by streams, have substantial lichen cover on boulders, and display significant soil development and tree growth (Fig. 3.3). They occur 0.1-3 km downstream from most Matthes deposits (Fig. 3.2), but commonly remain separated by many kilometers from the next older moraines, which are Tioga recessional deposits. As with the Matthes deposits, many of the Recess Peak glaciers in the southern Sierra Nevada appear to have had extensive covers of debris (Clark *et al.*, 1994a, Fig. 7). The insulation provided by the surficial debris on these glaciers enabled them to extend to lower elevations than equivalent bare-ice glaciers (Clark *et al.*, 1994a). However, their RW characteristics and stratigraphic position relative to Matthes deposits remain consistent, and they clearly correlate with deposits of relatively debris-free Recess Peak glaciers to the north.

ELA Estimates

The equilibrium line, which separates areas of net accumulation and ablation on glaciers, is often used as a proxy for alpine climatic snowlines, and thus climate, both modern and ancient. In the Sierra Nevada, regional ELA gradients for modern and late-Pleistocene glaciers rise approximately 12 to 16 m/km to the northeast across the range crest and about 2 m/km to the southeast parallel to the crest (Burbank, 1991; Gillespie, 1991).

The most common method of estimating ELAs of former glaciers, the accumulation-area ratio (AAR) method, relies on an empirical ratio derived from modern temperate glaciers that relates the former accumulation area to the total glacier area (e.g., Meierding, 1982; Leonard, 1984). The commonly used steady-state value for AARs of bare-ice glaciers is 0.65, although the ratio may be as small as 0.05 for extensively debris-covered glaciers (valley-floor rock glaciers; Clark *et al.*, 1994a; Chapter 2). For this portion of the study, I use an AAR of 0.65 for glaciers that were not extensively buried by debris, and an AAR of 0.15 for those that were (Clark *et al.*, 1994a). I distinguished the two types of deposits on the basis of extent and continuity of till (formerly the surficial rubble) upslope from the terminal moraines (Clark *et al.*, 1994a; Chapter 2).

The accuracy of ELA estimates using AARs depends on the former glaciers having mass balance regimes and area/altitude distributions similar to those of modern glaciers; however, errors are typically estimated as ± ~100 m (e.g., Meierding, 1982; Gillespie, 1991). Because of the small size and restricted altitudinal range of the glaciers in this study (typically less than 3 km in length, ~300-400 m altitudinal range), the accuracy using AARs in this study is probably closer to ± 50 m. Furthermore, the deposits I studied all lie within several kilometers of the Sierran crest, so that errors due to projection to a line that approximates the crest are within the error of the methods of estimation.

As discussed by Clark *et al.* (1994a), ELAs of small cirque glaciers such as those in this study strongly reflect local orographic effects on accumulation and ablation. ELAs of progressively smaller glaciers are increasingly influenced by these local effects. Burbank (1991) tried to minimize this effect by including only ELAs for glaciers that were at least 300 m wide. Clark *et al.* (1994a) demonstrated that even the glaciers that met this criterion remain strongly influenced by local topography, though typically less so than smaller glaciers. I therefore include even small glaciers in this study, although glaciers that are smaller than Burbank's 300-m width limit are distinguished in order to test this effect. Because local shading and accumulation patterns also depend on cirque aspect, only glaciers that had a northeasterly to northwesterly aspect are included, except where noted.

Sediment Coring

Because datable organic material is notoriously scarce in Sierran moraines, I focused on collecting and dating sediments from small lakes and bogs behind Recess Peak moraines that I had mapped and correlated as part of the ELA study above. AMS radiocarbon dating of small samples (< 1 g) of basal organic sediments from the cores provide high-precision minimum limiting ages for the Recess Peak advance. In this chapter, I present new radiocarbon dates from lakes formed inside Recess Peak ice limits. I also cored one bog that lies downstream from a Recess Peak moraine in an attempt to obtain a maximum limiting age on the advance, but I was unable to penetrate the apparent Recess Peak outwash layer beneath the bog.

I cored four widely separated sites that span much of the latitudinal extent of Recess Peak deposits in the Sierra Nevada (Fig. 3.1). North to south, the sites are: (1) a small unnamed tarn above Lake Aloha in the headwaters of the South Fork American River; (2) Green Treble Lake in Lee Vining Creek, immediately east of Yosemite National Park; (3) upper Garnet Lakes in the headwaters of Middle Fork San Joaquin River; and (4) Baboon Lakes on the Middle Fork, Bishop Creek. The cores comprise the highest suite of lake cores yet analyzed in the Sierra Nevada. I used a number of criteria in selecting the sites: (1) other than the exception mentioned above, sampled lakes all lie within or adjacent to the innermost Recess Peak moraines in that drainage; (2) the lakes are largely isolated from outwash of Matthes glaciers in the same drainages, except for the upper Garnet Lakes site; (3) upstream bedrock in the drainages is entirely granitic or other silicic rock and contain no significant sources of "old" carbon that could potentially affect radiocarbon dates; (4) the small volumes of the cored lakes and the high annual precipitation (typically > 170 cm/yr) in the basins means that residence times of water (and thus reservoir effects on carbon) in the lakes is on the order of years; and (5) maximum water depths are 4 m or less. Recess Peak and Matthes moraines at all sites are correlated on the basis of the mapping for the ELA study.

Cores were collected using a 2-inch (~5 cm)-diameter modified Livingston piston corer, operated from a portable floating plywood platform. The uppermost ~5 cm below the water-sediment interface typically was not recovered. Cores were extruded in the field, wrapped in plastic film and ABS casing, transported back to the lab, and stored in a cold room. In the lab, I split the cores from bottom-to-top to minimize effects of contamination by "young" carbon, and logged and photographed the sediments. Total organic carbon analyses on selected horizons, measured using the vapor-phase acidification method on a Carlo Erba CHN microanalyzer (Hedges and Stern, 1984), detected no significant inorganic carbon. Organic carbon percentages from the same analyses ranged from ~0.05% for basal outwash gravel to 16.25% for postglacial gyttja.

All dates in the cores are based on AMS-radiocarbon analyses performed at the Center for Accelerator Mass Spectrometry, Lawrence Livermore National Laboratory. Most dates are from small (< 1 g) samples of gyttja or organic silt, which span less than 1 cm of depth in the cores. Basal gyttja samples from the longest cores at each site were dated, as well as one or more other horizons higher up in all but one core to establish the validity of the basal dates. In addition, I dated adjacent samples of gyttja and macrofossils or peat and macrofossils in several cores to assess the accuracy of dating the peat or gyttja directly. All radiocarbon dates presented here are corrected for ^{13}C fractionation (Stuiver and Polach, 1977), and show 1σ analytic-error ranges. Calibrated radiocarbon ages (cal yr B.P.), corrected for variations in atmospheric ^{14}C content (Bard *et al.*, 1990; Stuiver *et al.*, 1991) were calculated using the calibration program CALIB (v. 3.0.3b; Stuiver and Reimer, 1993), with age ranges reflecting combined 1σ analytic and calibration-curve standard errors. In this paper, I present only the radiocarbon dates and basic stratigraphies of the cores. Detailed sedimentological and palynological analyses of the cores are underway at the Desert Research Institute, Reno, Nevada.

RESULTS

ELA Trends

ELA estimates of Recess Peak and Matthes glaciers along the Sierran crest exhibit large but systematic variability (Fig. 3.5). The regional trend shows ELAs declining to the north, as would be anticipated from modern precipitation and temperature gradients (Dale, 1966). The trend is not monotonic, however; substantial local variability at several different scales is superimposed on the broad latitudinal trend. Major negative deviations coincide with the headwaters of major river gorges on the western slope (e.g., the San

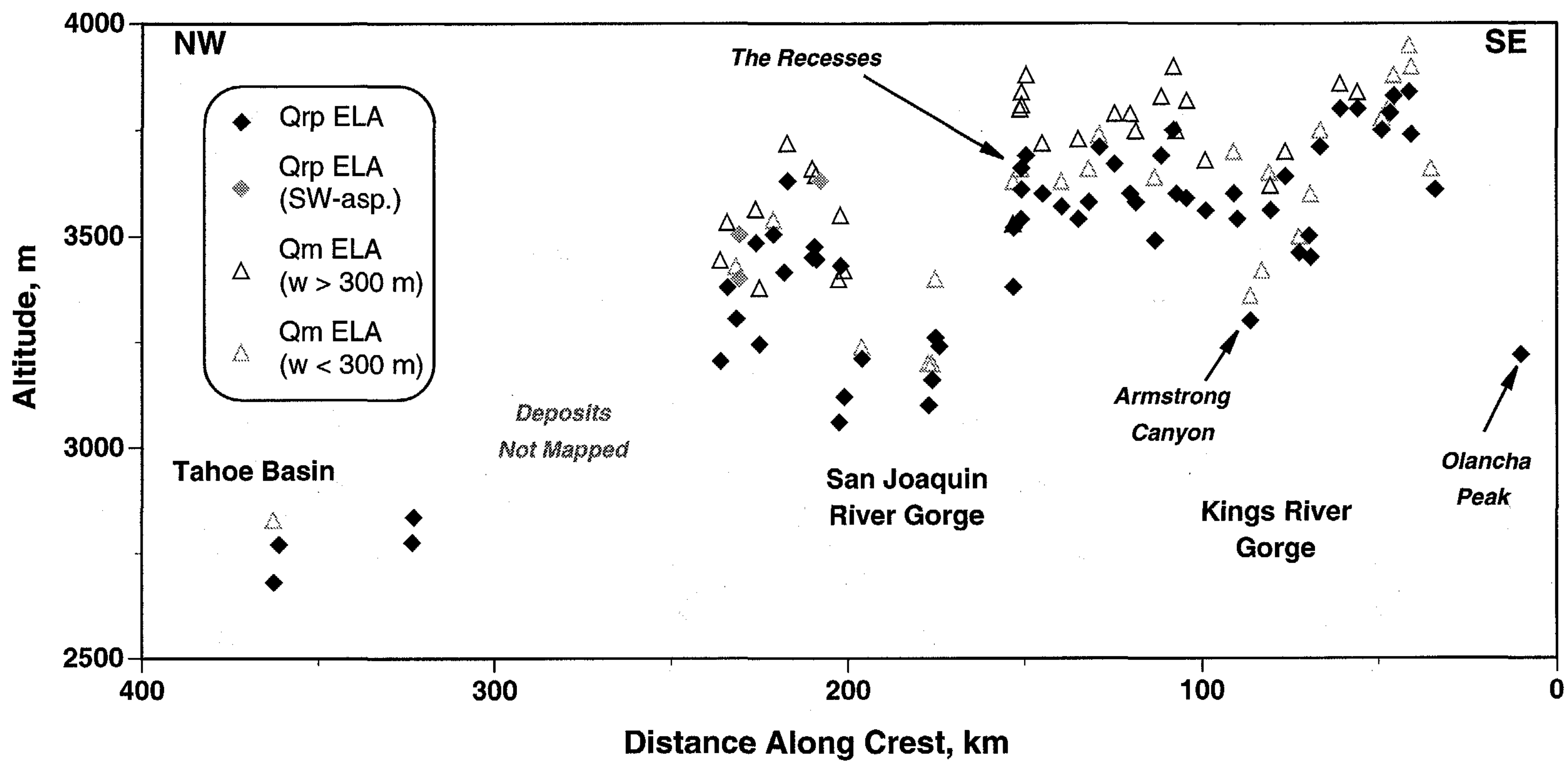


Figure 3.5: NW-SE profile of reconstructed Recess Peak (Qrp) and Matthes (Qm) ELAs along crest of Sierra Nevada from Olancha Peak to Lake Tahoe. ELAs calculated and plotted following Clark *et al.* (1994a). All ELAs are for glaciers with northerly aspects, except for three Recess Peak ELAs with southwestern aspects (gray diamonds) north of San Joaquin River Gorge. Matthes ELAs are divided into those for glaciers having widths (w) greater than 300 m (black triangles) and those with w < 300 m (gray triangles). All reconstructed Recess Peak glaciers have w > 300 m, and are no more than 3 km from line of profile. Compare with Figure 2.9.

Joaquin and Kings river gorges) whereas intervening major positive deviations lie east of secondary drainage divides west of the range crest (e.g., Great Western, Monarch, and Goddard divides, Cathedral Range; Fig. 3.1). The negative deviations probably reflect funneling of Pacific storms along major NE-trending drainage troughs, allowing greater precipitation to reach further into the range, thereby lowering ELAs; conversely, the positive deviations reflect rainshadow effects of high divides in the upwind (southwest) direction (e.g., Porter, 1975b, p. 35; Burbank, 1991, p. 299). As predicted in the earlier paper (Clark *et al.*, 1994a, p. 151), the largest apparent deviation coincides with the headwaters of the San Joaquin River gorge, which forms the largest topographic gap in the southern half of the range (Figs. 3.1, 3.5).

Within these major deviations, ELAs vary on a more local scale. These local fluctuations mimic topographic variations along the adjacent crest, and apparently reflect local orographic effects, such as shading, avalanching, and wind-drifting, which increase accumulation and reduce ablation at those sites (Clark *et al.*, 1994a, Fig. 8). The most pronounced examples of these influences occur in the unusually deep cirques at Armstrong Canyon, opposite the headwaters of the Kings River gorge, and off the northeastern slope of Olancha Peak at the southern end of the glaciated range (Fig. 3.5). The unusually low altitudes of glaciers at both sites, however, also probably reflect regional topographic influences on accumulation patterns as well: storm funneling in the Kings River gorge for the former, absence of significant topographic divides to the southwest of Olancha Peak for the latter. It is also possible that, because there is no Matthes deposit against which to compare it, the putative Recess Peak deposit at Olancha Peak is instead related to an earlier Tioga recession. This seems unlikely, however, because of the distinctive morphologies and spatial separation (several kilometers) between the Recess Peak and Tioga moraines.

Together, these orographic influences locally overwhelm the regional latitudinal trends of the Matthes and Recess Peak ELAs. This finding underscores the concept that the ELAs of small cirque glaciers generally reflect the orographic snowline, rather than the climatic snowline (e.g., Matthes, 1942; Meierding, 1982; Clark *et al.*, 1994a). The influence of these local factors was smaller when the glaciers were at their maximum late-Pleistocene positions (Burbank, 1991; Gillespie, 1991). The difference may be 1) that large glaciers partially filled many of the canyons on the western slope, blocking the storm funneling; and 2) that accumulation zones reached well beyond the conserving confines of the cirques, thus allowing the climatic snowline to overwhelm the orographic snowline (Clark *et al.*, 1994a).

However, despite their irregularity, Recess Peak ELAs generally vary coherently with Matthes ELAs, typically remaining 100-200 m lower than Matthes ELAs in the same cirques (Figs. 3.5, 3.6). Similarities in the gross trends are especially apparent (Fig. 3.5), although differences between the ELAs (ΔELAs) show substantial variability (Fig. 3.6). The differences, however, are less than the variability of the regional trends (up to 400-500 m). Differences in hypsometries of some cirques accounts for some of the variability in ELAs between the two advances, as in the case of the glaciers near Banner Peak (Fig. 3.6). The steep slopes there apparently permitted the Recess Peak glaciers to have ELAs nearly 300 m lower than those of the much smaller Matthes glaciers (Fig. 3.6; see also Fig. 3.10 in the section on coring). In contrast, there seems to be smaller separations between the ELAs of the two advances at the southern end of the range. Apparently, glaciers of both advances were restricted to the cirque headwalls because of greater dependence on local factors such as shading, accumulation, and debris cover in this more rugged part of the range. When local factors such as those discussed above are considered, however, the ΔELAs remain relatively consistent, especially considering that they can vary by more than 100 m in a single drainage. Moreover, if the two outliers at

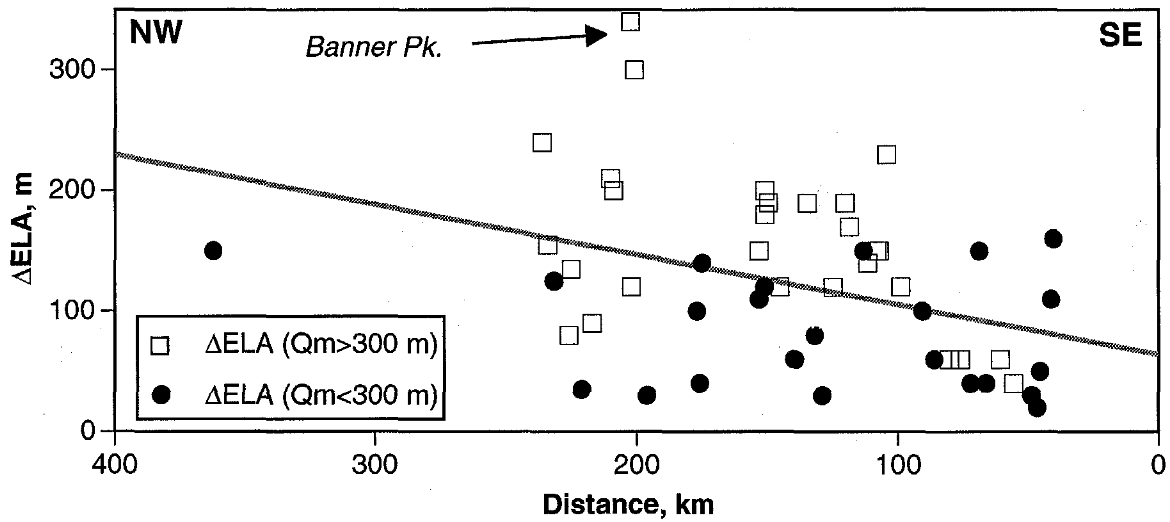


Figure 3.6: Difference in ELAs (ΔELA) from reconstructed Matthes and Recess Peak glaciers shown in Figure 3.5. ΔELAs separated according to maximum width (w) of Matthes (Qm) glaciers (Figure 3.5). Line is regression to all data ($r2 = 0.15$). Variability reflects strong local orographic effects on ELAs for these small glaciers. Slope of line primarily reflects large ΔELAs for former glaciers at Banner Peak, and smaller ΔELAs towards southern end of range where both Recess Peak and Matthes glaciers dependended increasingly on local orographic shading and accumulation and thus were both restricted to heads of cirques.

Banner Peak are ignored, the ΔELA-trend is relatively flat with distance along the crest. The similar ELA trends, in combination with the spatial distribution and geomorphic character of the moraines, help confirm my correlations of the two advances from the type area (Birman, 1964). They also demonstrate that the spatial fluctuations of the ELAs for both advances are valid, as also shown by Clark *et al.* (1994a) for a smaller sample.

Recess Peak glaciers in south-facing cirques were rare and generally much smaller than those in adjacent north-facing cirques. I estimated ELAs of three such glaciers in order to assess the affect of aspect on ELAs (Fig. 3.5). Although the sample is small, it suggests that southern aspects raise ELAs by ~50-200 m. Because the effects of shading in south-facing cirques are far less than in north-facing cirques, glaciers in the former may more closely represent the climatic snowline. I observed no deposits of Matthes age with southern aspects.

Sediment Cores

I collected several cores from each of the four sites in order to confirm the general post-Recess Peak stratigraphy at each site. The cores range in length from 15 cm to 212 cm (measured in the lab) and are dominated by dark brown, generally massive gyttja; all cores except those from upper Garnet Lakes are from the deepest parts of the lakes. Only cores from the southernmost site (Baboon Lakes) bottomed in reduced, inorganic outwash; all others bottomed in gyttja or tephra some distance above the basal (Recess Peak) outwash. Because all sites except upper Garnet Lakes are isolated from Matthes outwash, I did not observe nor expect Little Ice Age inorganic sediment spikes. Compression of sediments in the cores during coring was typically 10-15%. Descriptions of AMS-radiocarbon dates from all cores related to the Recess Peak advance are listed in Table 3.1.

Table 3.1: Radiocarbon Ages from Recess Peak Cores, Sierra Nevada

Sample Name	Location	Material	Core #	Depth, cm	^{14}C yr B.P. (1σ error)[§]	cal. yr B.P. (1σ)*
GT-C-1	Green Treble Lk., Hall Wilderness	basal organic silt	GTL-4	115-116	7750 ± 60	8550-8415
GT-C-2	Green Treble Lk., Hall Wilderness	basal gyttja above highest basal organic silt	GTL-4	101-102	5870 ± 60	6750-6645
GT-C-3	Green Treble Lk., Hall Wilderness	gyttja, base of lower tephra	GTL-4	91-92	4630 ± 50	5330-5300
GT-C-4	Green Treble Lk., Hall Wilderness	gyttja, base of upper tephra	GTL-4	28-29	2080 ± 50	1970-1960
DW-C-1a	Desolation Wilderness tarn	charcoal, top of lower tephra	DW-5	16-17	7130 ± 230	8130-7670
DW-C-1b	Desolation Wilderness tarn	organic silt next to DW-C-1a, top of lower tephra	DW-5	16-17	7140 ± 70	7960-7895 7855-7840
DW-C-2	Desolation Wilderness tarn	gyttja, top of upper tephra	DW-5	7	6740 ± 60	7585-7530 7490-7480
BL-C-1	upper Baboon Lks., Bishop Cr.	organic silt, near top of gyttja/outwash transition	BL-2	115-116	11,190 ± 70	13,185-13,015
BL-C-2	upper Baboon Lks., Bishop Cr.	charcoal in gyttja	BL-2	106.5	9450 ± 50	10,780-10,765 10,545-10,365
BL-C-3	upper Baboon Lks., Bishop Cr.	chip from large branch	BL-2	95.5	8490 ± 60	9495-9435
BL-C-4	upper Baboon Lks., Bishop Cr.	gray-green gyttja, next to BL-C-3	BL-2	95-96	9360 ± 70	10,390-10,290 10,250-10,220
BL-C-5	upper Baboon Lks., Bishop Cr.	wood chips below upper clayey silt	BL-2	40	7050 ± 70	7915-7755
BL-C-6	Baboon Lks. bog	peaty silt above basal sandy unit	BL-3	151-152	9040 ± 110	10,045-9925

(continued on next page)

Table 3.1 (continued).

Sample Name	Location	Material	Core #	Depth, cm	^{14}C yr B.P. (1σ error)§	cal. yr B.P. (1σ)*
BL-C-7	Baboon Lks. bog	wood chip	BL-3	139.5	6950 ± 50	7790-7665
BL-C-8	Baboon Lks. bog	peat, next to BL-C-7	BL-3	140	7020 ± 60	7900-7855 7845-7720
BL-C-9	Baboon Lks. bog	top of peat below middle sandy unit	BL-3	114-115	5710 ± 90	6640-6410
BL-C-10	Baboon Lks. bog	basal peat above middle sandy unit	BL-3	83	5580 ± 90	6450-6290
BL-C-11	upper Baboon Lks., Bishop Cr.	twig immediately above thin tephra	BL-2	83	7900 ± 60	8945-8905 8725-8555
BL-C-12	upper Baboon Lks., Bishop Cr.	gyttja next to wood in BL-C-11	BL-2	83	7890 ± 60	8945-8910 8725-8550
BL-C-13c	central Baboon Lks., Bishop Cr.	charcoal, base of dark green gyttja	BL-1	150-151	10,170 ± 170	12,145-11,425
BL-C-13g	central Baboon Lks., Bishop Cr.	gyttja, next to BL-C-13c	BL-1	150-151	10,120 ± 80	11,955-11,430
BL-C-14	central Baboon Lks., Bishop Cr.	gyttja, top of gyttja/outwash transition	BL-1	160-161	10,880 ± 60	12,880-12,725
UGL-C-1	upper Garnet Lks., Ritter Range	basal gyttja	UGL-2	158	3420 ± 50	3620-3700

§All samples analyzed at CAMS, Lawrence Livermore National Laboratory

*(Stuiver and Reimer, 1993)

South Fork American River. The northernmost coring site is a small unnamed tarn formed inside lateral and terminal moraines of a Recess Peak glacier above Lake Aloha (Fig. 3.7). The moraines in this drainage were initially mapped by M. Clark (1973, *unpublished data*) and Whiting (1985, 1986). Two cores from the tarn were the shortest of any cores in this study (Fig. 3.8A). The cores bottomed in a thick, fine-grained, gray tephra initially encountered at about 25 cm depth. A thinner, coarser tephra unit occurs at about 19-21 cm depth. Liquifaction of the lower tephra, which was at least 25 cm thick, prevented coring through it. Radiocarbon dates of charcoal and organic sediments directly overlying each of these layers demonstrate that the tephra units are related to the eruption of Mt. Mazama (ancestral Crater Lake, Oregon)(Table 3.1). Dates of 7130 ± 230 and 7140 ± 70 ^{14}C yr B.P. (~7900 cal yr B.P.) for charcoal and organic silt, respectively, from a single horizon immediately above the lower tephra indicate that it may be related to the Tsoyawata eruption, a precursor event to the culminating Mazama eruption (Bacon, 1983). The date of 6740 ± 60 ^{14}C yr B.P. (~7500 cal yr B.P.) for gyttja overlying the upper tephra indicates that it may represent the culminating eruption (Sarna-Wojcicki *et al.*, 1991). Further petrographic and geochemical analyses of the tephras may help confirm this assessment. Adam (1967) and Edlund (1993) found Mazama tephra in cores of a nearby bog and lake, respectively, dammed behind Tioga recessional moraines, supporting my inferences on the tephras.

There is no sedimentological evidence for any post-Recess Peak glacier advances in the cores. The post-Mazama deposits are very thin, however, and suggest that deposition may have slowed or ceased for a period during the latter half of the Holocene, perhaps as a result of late Holocene desiccation of the shallow (~2 m) pond (e.g., Stine, 1994). Unconformities are not apparent, so substantial erosion is unlikely. Whiting (1985, 1986) divided moraines adjacent to the pond into two ages, late Pleistocene and Neoglacial, primarily on the basis of relative weathering (RW) data. Whiting's (1985,

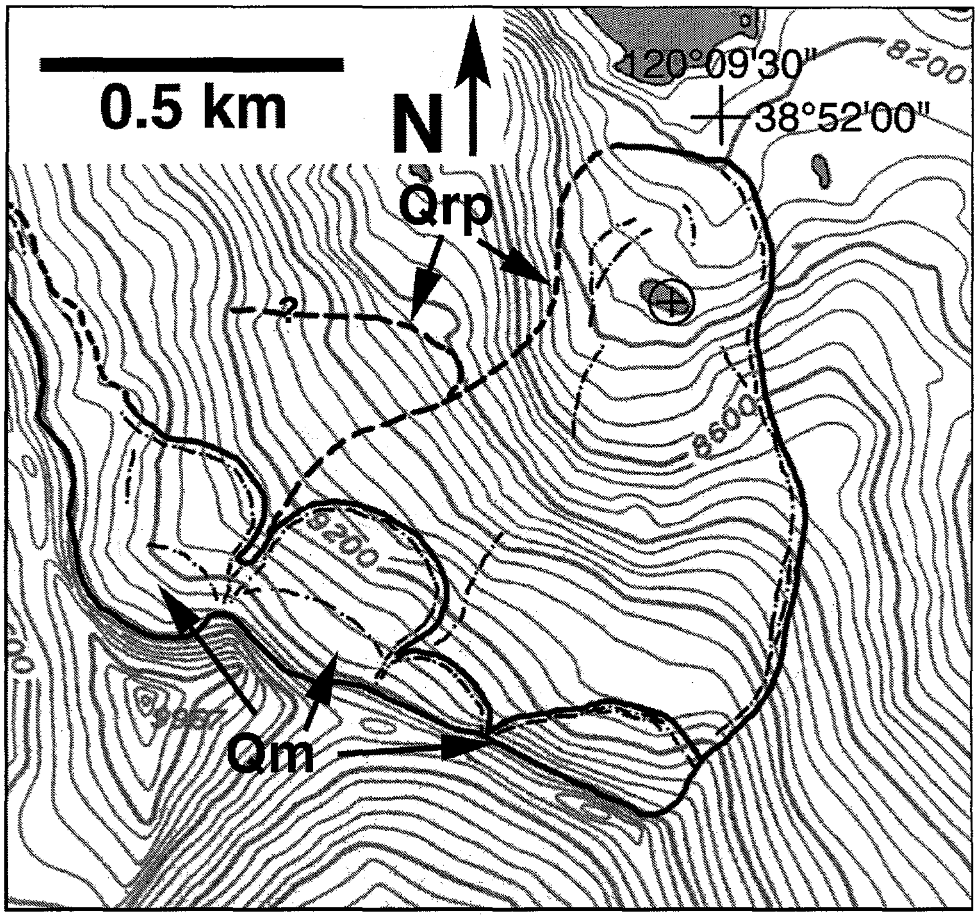
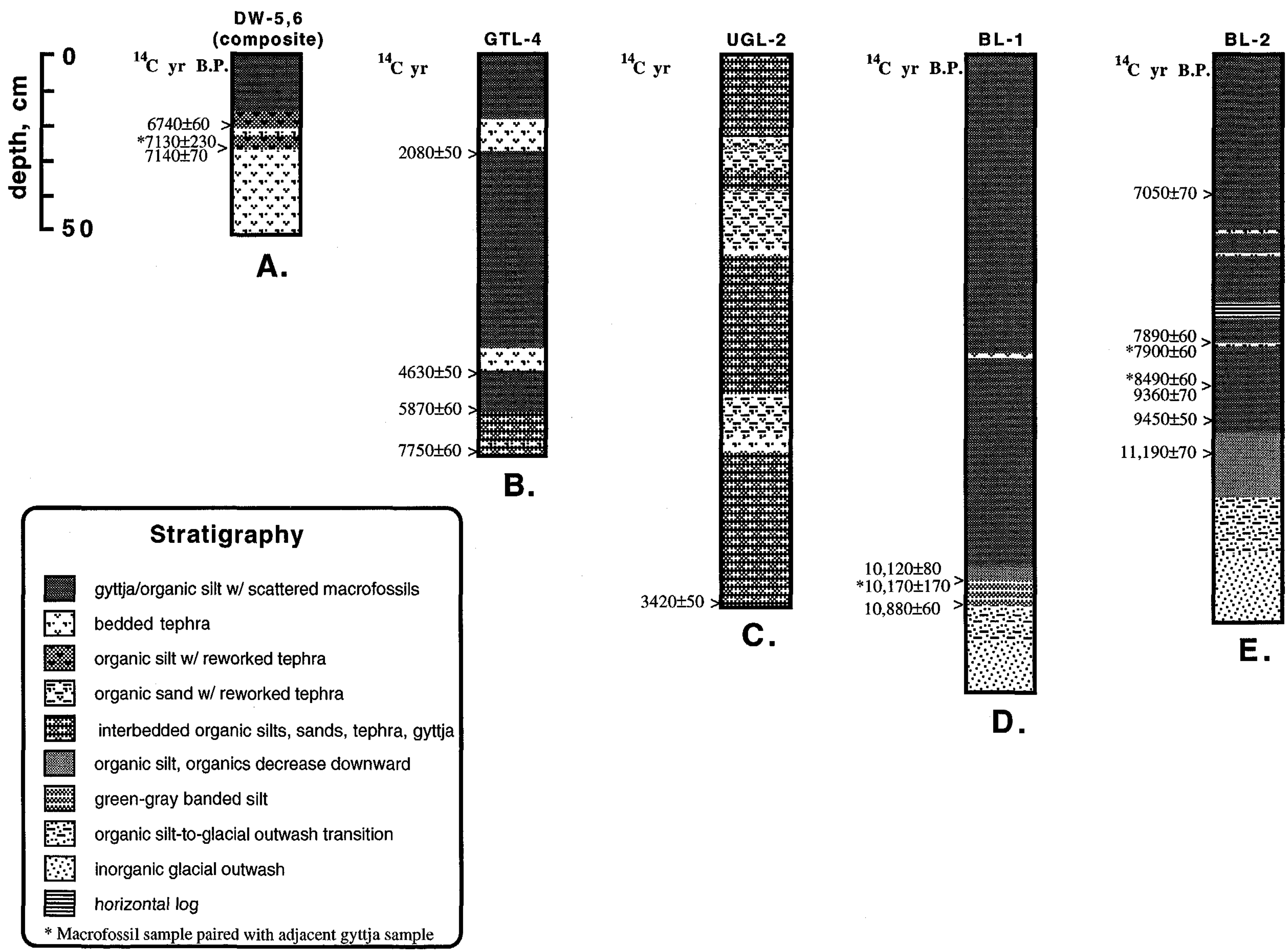


Figure 3.7: Map of Matthes (Qm) and Recess Peak (Qrp) moraines and ice limits near Lake Tahoe basin. Coring site in small tarn near Recess Peak terminus shown by circle-and-crosshair. Moraine crests shown as dash-dot lines; former glacier limits shown by bold lines, dashed where approximate, queried where uncertain. Lake Aloha shown along upper edge of figure. Topography from U.S. Geological Survey Pyramid Peak 7.5' quadrangle, 40-ft. (12.2-m) contours.

Figure 3.8: Simplified stratigraphies of cores for (A) tarn above Lake Aloha, (B) Green Treble Lake, (C) upper Garnet Lakes, (D, E) Baboon Lakes, in the Sierra Nevada. Core A is a composite of two cores. Locations and ^{14}C ages (with 1σ errors) of AMS radiocarbon samples are shown to the left of each column. Paired ages at single location indicate adjacent macrofossils and gyttja samples. These dates and others from cores not shown are listed in Table 3.1. Cores from Baboon Lakes (D, E) terminate in outwash.



depth, cm
0
50
DW-5,6 (composite)
^{14}C yr B.P.
6740±60
*7130±230
7140±70
A.
GTL-4
^{14}C yr
2080±50
4630±50
5870±60
7750±60
B.
UGL-2
^{14}C yr
3420±50
C.
BL-1
^{14}C yr B.P.
10,120±80
*10,170±170
10,880±60
D.
BL-2
^{14}C yr B.P.
7050±70
7890±60
*7900±60
*8490±60
9360±70
9450±50
11,190±70
E.
Stratigraphy
gyttja/organic silt w/ scattered macrofossils
bedded tephra
organic silt w/ reworked tephra
organic sand w/ reworked tephra
interbedded organic silts, sands, tephra, gyttja
organic silt, organics decrease downward
green-gray banded silt
organic silt-to-glacial outwash transition
inorganic glacial outwash
horizontal log
* Macrofossil sample paired with adjacent gyttja sample

Table 7) data, however, are equivocal: he had four sampling sites on his "late-Pleistocene" moraines, and only two directly adjacent sites on a single short "Neoglacial" moraine. The primary RW distinction between the moraines is that boulders on the former moraines are apparently more rounded than in the latter; differences in other RW parameters (e.g., weathering rinds, fresh-to-weathered ratios, moraine morphologies) are either inconsistent or statistically insignificant. However, striae and polish on the boulders from the "late-Pleistocene" moraines clearly indicate that they were transported and rounded subglacially, whereas those on the putative "Neoglacial" moraine are not polished or striated. The small size, uniform lithology, and angularity of the latter instead indicate that they formed as a supraglacial rockfall deposit. The lack of rounding in this deposit is therefore probably original and not a result of a "young" age, especially considering the similarity of the other RW measurements. I thus concur with M. Clark's (1973, *unpublished data*) inference based on mapping that the moraines below the small Matthes moraines all belong to a single Recess Peak advance. Bedrock and moraine geometries (Fig. 3.7) dictate that any Neoglacial advance as large as that proposed by Whiting (which would have terminated at or in the pond) would have deposited substantial outwash in the pond. The absence of any significant post-Mazama clastic sediments in the pond or any moraines between the Matthes and Recess Peak deposits demonstrates that the Matthes advance was the most extensive of the late Holocene in this area.

Lee Vining Creek. Green Treble Lake is dammed by the innermost Recess Peak terminal moraine of a glacier that flowed off the northeastern ridge of White Mountain, in the headwaters of Lee Vining Creek (Fig. 3.9). One of the sample sites used by Nishiizumi *et al.* (1989) to calibrate *in-situ* production rates of cosmogenic ^{10}Be and ^{26}Al lies directly downstream from similar Recess Peak deposits in the adjoining Conness Lakes drainage 3 km north of Green Treble Lake.

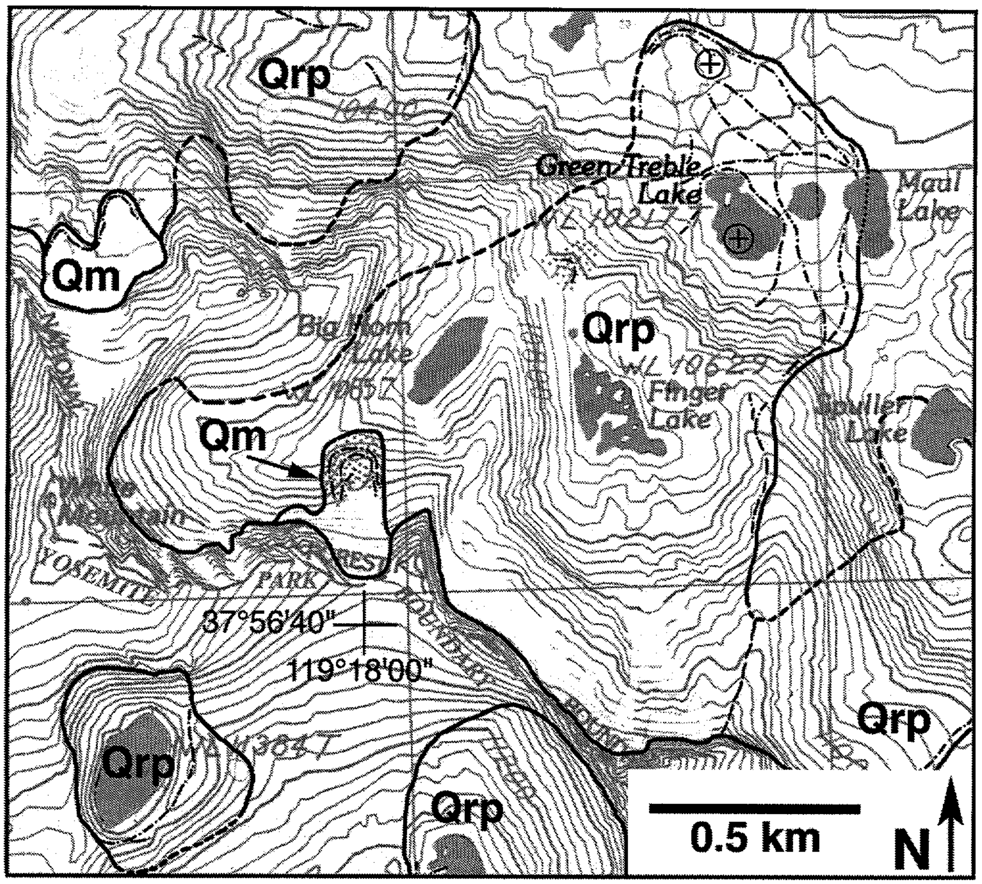


Figure 3.9: Map of Matthes (Qm) and Recess Peak (Qrp) moraines and ice limits on eastern slope of White Mountain in the headwaters of Lee Vining Creek. Coring sites, including Green Treble Lake, shown by circle-and-crosshair in upper right. Shallow core from meadow north of Green Treble Lake not analyzed. Symbols as in Figure 3.7. Sierran crest forms Yosemite National Park - Inyo National Forest boundary. Topography from U.S. Geological Survey Tioga Pass 7.5' quadrangle, 40-ft. (12.2-m) contours.

A core from the deepest part of the lake bottomed in laminated greenish organic silt at roughly 175 cm depth, with 126 cm recovered (Fig. 3.8B). Hole collapse and friction prevented us from coring deeper. Sediments in the upper part of the core consist primarily of massive greenish-brown gyttja with scattered sand grains and thin sand layers. Two thick, fine-grained, well bedded, light-gray tephras occur at 28-38 and 94-101 cm depth. The tephras are uncontaminated and have sharp upper and lower contacts. A date of 2080 ± 50 ^{14}C yr B.P. (1960-1970 cal yr B.P.) for gyttja underlying the younger tephra indicates that it correlates with a tephra occurring at similar depths and overlying gyttja dated 1910 ± 80 ^{14}C yr B.P. (1730-1930 cal yr B.P.) in a pond at Tioga Pass, 5 km southeast of Green Treble Lake (Anderson, 1990). The older tephra may correlate with two thin tephras in the Tioga Pass pond core that overlie gyttja dated to 4060 ± 160 ^{14}C yr B.P. (4310-4830 cal yr B.P.)(Anderson, 1990). However, gyttja underlying the lower Green Treble Lake tephra yielded a date of 4630 ± 50 ^{14}C yr B.P. (5300-5330 cal yr B.P.), suggesting that the tephra is from a separate, earlier eruption. Anderson concluded that the tephras in his cores are related to eruptions in the Mono and Inyo craters to the southeast. The basal 14 cm of the Green Treble Lake core consists of interbedded brownish-green organic sands and silts. Because I did not reach outwash in this core, the age of the basal sediments (7750 ± 60 ^{14}C yr B.P.; 8410-8550 cal yr B.P.) does not closely limit the time of retreat of the Recess Peak glacier.

Middle Fork San Joaquin River. A set of small, shallow lakes on a bench above Garnet Lake, informally referred to here as upper Garnet Lakes, is formed behind Recess Peak moraines from a glacier that flowed down the northeastern slope of Banner Peak (Fig. 3.10). The uppermost and largest of these lakes lies within the innermost distinct Recess Peak moraine. The receding remnant of a small Matthes glacier occupies the head of the cirque. Other well developed Recess Peak and Matthes moraines occupy the upper end of Thousand Island Lake and the other slopes of Banner Peak (Fig. 3.10).

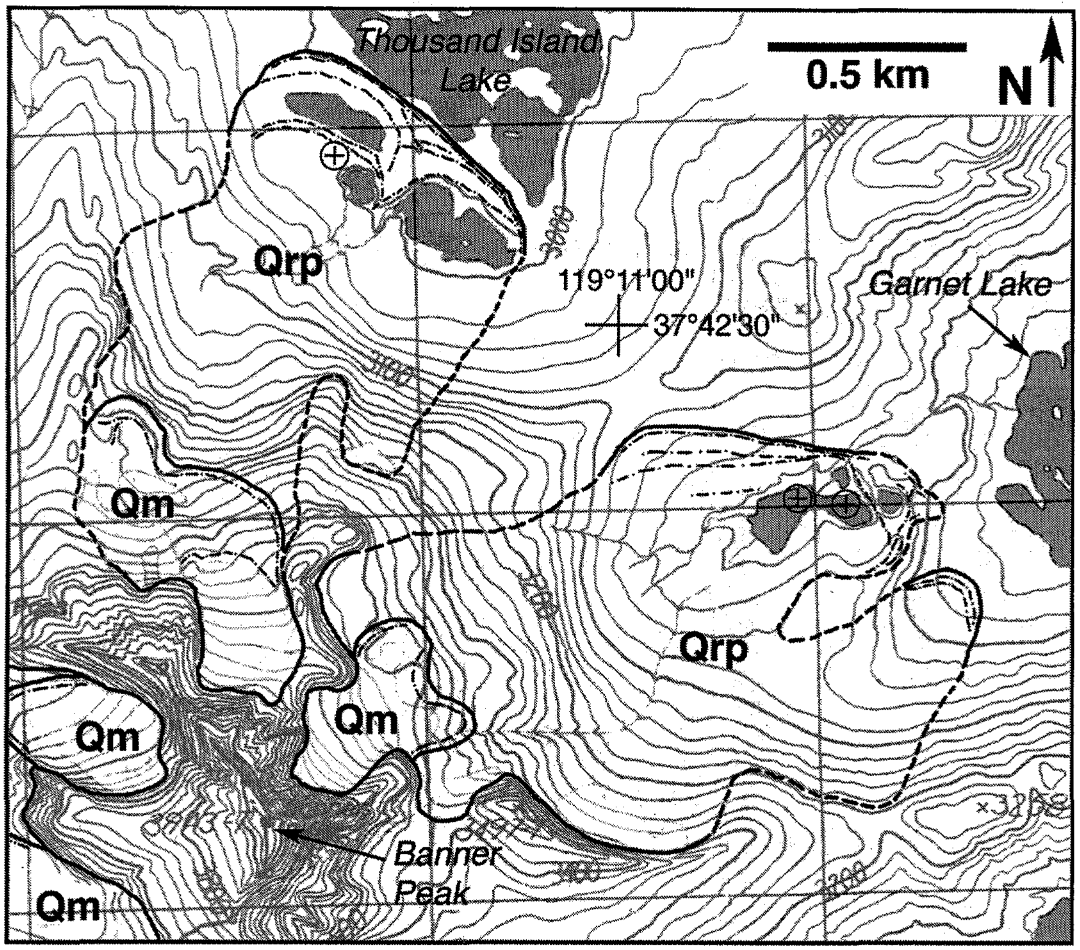


Figure 3.10: Map of Matthes (Qm) and Recess Peak (Qrp) moraines and ice limits next to Banner Peak, headwaters of the Middle Fork San Joaquin River. Coring sites shown by circle-and-crosshair. Shallow core from meadow behind Thousand Island Lake Qrp moraine not analyzed. Symbols as in Figure 3.7. Topography from U.S. Geological Survey Mt. Ritter 7.5' quadrangle, 20-m. contours.

The presence of gravel layers prevented cores longer than ~30 cm from behind these moraines, however.

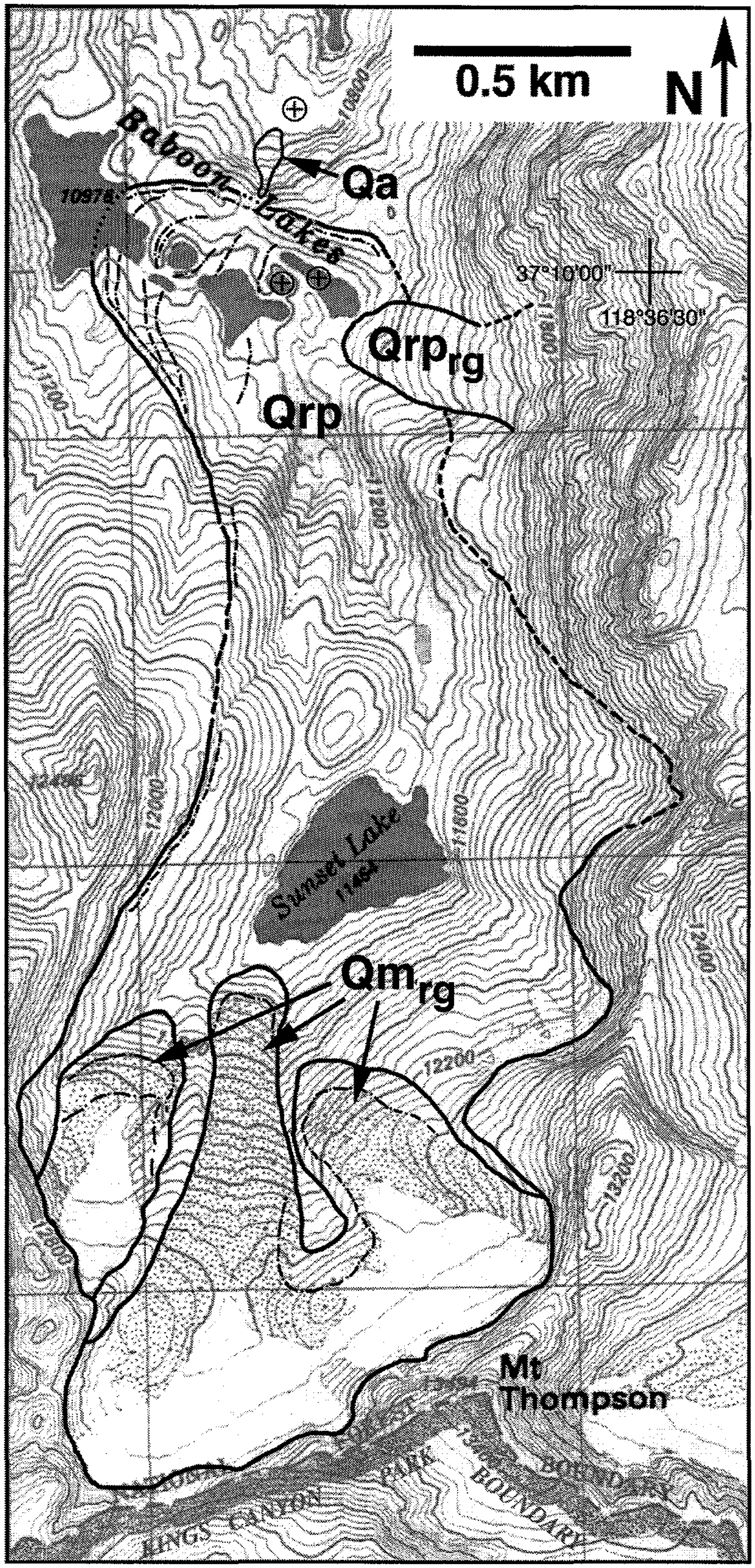
I collected cores in water depths of ~1 m in each of the two sub-basins that comprise the largest upper Garnet lake. Both reached sediment depths of about 160 cm; friction and collapse of the holes between successive pushes prevented deeper cores. Stratigraphy in both cores is complex, consisting of interbedded sands, organic silts, and tephras (Fig. 3.8C). Reworked tephras are ubiquitous in the cores. Oxidized horizons in the upper part of the cores suggest that the shallow lakes have been periodically exposed subaerially in the late Holocene. Neither core encountered inorganic outwash, and coarseness of the sediments suggest relatively high sedimentation rates in these lakes. The basal age of core UGL-2, 3420 ± 50 ^{14}C yr B.P. (3620-3700 cal yr B.P.), also supports relatively high sedimentation rates. The young basal age and high sedimentation rate indicate that the basal age does not closely date the retreat of the Recess Peak glaciers in this area.

Although this location is fed by meltwater from a small Matthes glacier (unlike the other coring sites), there are no obvious changes in sedimentation in the cores to indicate the onset of Neoglaciation. However, the high clastic sedimentation rate in the lakes, largely resulting from the abundant Holocene tephra fall in the basin, probably obscures any changes in sedimentation that may have accompanied the formation of the Matthes glacier. Furthermore, considering the small size of the glacier, I would not expect a large change in sedimentation during its formation. It is possible, however, that Neoglaciation in the drainage began before ~3700 cal yr B.P. (i.e., pre-Little Ice Age), although previous palynologic, dendrochronologic, and isotopic work elsewhere in the region (e.g., Adam, 1967; LaMarche, 1973; Scuderi, 1984; 1987a; Anderson, 1990; Feng and Epstein, 1994) suggests this is unlikely. In any case, any such advance would have been even smaller than the Matthes.

Bishop Creek. Baboon Lakes (Fig. 3.1) comprise a suite of small- to medium-sized lakes formed behind a bedrock sill that crosses a tributary of Middle Fork Bishop Creek (Fig. 3.11). A Recess Peak glacier originating on the northwest slope of Mt. Thompson terminated against this buttress and flowed into the lowest and largest lake. Three active Matthes debris-covered glaciers occupy the headwaters of the drainage. There are no moraines between those of the Recess Peak and Matthes advances. I cored two small lakes that are separated from the main drainage by till-mantled bedrock inside the innermost Recess Peak recessional moraine (Fig. 3.11). The lakes are thus isolated from outwash of Matthes and modern glaciers in the headwaters of the creek. A small, inactive valley-wall rock glacier terminates near the upstream edge of the upper lake I cored. The rock glacier appears to have flowed for a short time after retreat of the Recess Peak glacier, since it cross-cuts the Recess Peak lateral moraine; its association with the Recess Peak advance, however, is indicated by its inactivity and morphologic and weathering similarities to the adjacent Recess Peak moraines. It may also imply a cool but dry period immediately following the retreat of the Recess Peak glacier.

The cores from both lakes, taken from water depths of 2-3 m, are about 180 cm long and dominated by massive dark greenish-brown gyttja, with intermittent slightly lighter and siltier intervals and scattered macrofossils (Fig. 3.8D, E). The cores also contain several very thin, indistinct tephras that probably correlate with Holocene eruptions of Mono or Inyo craters (Anderson, 1990). The bottom 10-15 cm of both cores consists of reduced, inorganic silty sand that I interpret as Recess Peak outwash. The transition from inorganic outwash to gyttja occurs over about 30-40 cm in the cores, although there is a relatively distinct light-grayish-green organic silt unit within the transition in core BL-1 (Fig. 3.8D). Dates from the upper part of the outwash-gyttja transition in these cores are 10,880 ± 60 ^{14}C yr B.P. (~12,800 cal yr B.P.) for BL-1 and 11,190 ± 70 ^{14}C yr B.P. (~13,100 cal yr B.P.) for BL-2. The older of these dates provides

Figure 3.11: Map of Matthes (Qm) and Recess Peak (Qrp) moraines and ice limits on northwest slope of Mt. Thompson in headwaters of Middle Fork Bishop Creek. Coring sites shown by circle-and-crosshair. Symbols as in Figure 3.7. Qm_{rg} and Qrp_{rg} denote rock glacier deposits of respective advances; former are still active, latter probably formed at end of Recess Peak advance and is now extinct. Qa indicates avalanche debris originating from Recess Peak right-lateral moraine. Sierran crest forms Kings Canyon National Park - Inyo National Forest boundary. Topography from U.S. Geological Survey Mt. Thompson, California 7.5' quadrangle, 40-ft. (12.2-m) contours.



0.5 km
N
Baboon Lakes
Qa
10978
37°10'00"
118°36'30"
Qrp_rg
Qrp
Sunset Lake
11464
Qm_rg
12200
Mt Thompson
KINGS CANYON PARK BOUNDARY

the closest minimum age estimate yet determined for the end of the Recess Peak stade in the Sierra Nevada. The absence of oxidized layers or unconformities indicates that the lakes, although shallow, have not desiccated during the Holocene.

I cored a small bog immediately downstream from the right-lateral moraine in an attempt to obtain a maximum date for the Recess Peak advance, but was unable to penetrate a coarse sand layer, possibly from the Recess Peak glacier. The radiocarbon age of the basal peat in this core (BL-3; Table 3.1), 9040 ± 110 ^{14}C yr B.P. (~10,000 cal yr B.P.), does not appear to be a close limiting minimum age for the Recess Peak advance because it is younger than the basal ages of the cores from inside the moraines. It may instead represent a period of moraine instability after the glacier receded as suggested by a gap in the right-lateral moraine above a talus that terminates in the bog.

The validity of the oldest dates in the Baboon Lakes cores is supported by many other dates from overlying sediments in each core (Fig. 3.8D, E). Dates of adjacent macrofossils and gyttja (Table 3.1) at 83-cm depth in core BL-2 (samples BL-C-11, -12) and 150 cm in core BL-1 (samples BL-C-13c, -13g), and adjacent peat and a macrofossil at 140 cm in BL-3 (samples BL-C-7, -8), are indistinguishable within one standard deviation. Similar results from adjacent samples were obtained from other cores as well (e.g., charcoal and organic silt in samples DW-C-1a, -1b, Table 3.1; Fig. 3.8A), indicating stable carbon in all cores. A sample of a large branch at 95 cm depth in core BL-2 (sample BL-C-3) was ~1000 yr younger than adjacent gyttja (sample BL-C-4); considering the similar ages of the adjacent macrofossils and gyttja mentioned above, and the slow sedimentation rates in the lakes, it seems likely that this large branch sank into older flocculent gyttja when it was deposited. All other paired samples involve small macrofossils (e.g., small twigs) that would not have been affected by this problem.

In summary, my coring data from four widely separated areas along the crest of the Sierra Nevada support a late-Pleistocene age for the Recess Peak advance. Cores from the southernmost site, Baboon Lakes, are the only ones that penetrated to outwash, and therefore give the closest minimum dates for the advance. Cores from the other sites bottomed at stratigraphically higher horizons and therefore do not provide close limiting ages for the advance. However, none supports a Neoglacial age for the advance, and my regional mapping and ELA estimates strongly support my correlations of Recess Peak deposits between coring sites.

DISCUSSION

Re-evaluation of the Sierran glacial sequence indicates that the Recess Peak advance was a late-Pleistocene event and was the most extensive regional glacier advance following the widespread retreat of the Tioga glaciers. Geomorphic mapping and sediment cores from alpine lakes and meadows in the central Sierra Nevada indicate that no significant expansion of glaciers (e.g., the "Hilgard glaciation" of Birman (1964)) occurred between the Tioga maximum and Recess Peak advances (Clark, 1976; Clark and Clark, 1995). Spatial separation and differences of thousands of years in oldest minimum ages for Tioga and Recess Peak deglaciations (D. Clark, unpublished data) suggest that the Recess Peak was a separate advance and not merely a Tioga recessional event. However, in the absence of direct stratigraphic evidence, it is also possible that Recess Peak deposits represent the last readvance of retreating Tioga ice.

Stratigraphic and morphologic affinities of Tioga recessional moraines with Birman's (1964) adjacent type Hilgard moraines in Mono and Rock creeks show that the "Hilgard" deposits are actually late-stage Tioga medial and recessional moraines. Moreover, recent reevaluations (Clark and Clark, 1995; M. Clark, 1995, personal communication) of the boulder-weathering ratios that Birman relied on to distinguish the type Hilgard moraines from adjacent Tioga recessional moraines failed to find such a distinction. The moraines instead appear to record the last major stillstand or readvance of the retreating Tioga glaciers before they stagnated and likely disappeared (Clark, 1976; Clark and Clark, 1995). My mapping further supports Clark's (1976) contention that there is no substantive evidence for a separate stade between the Tioga and Recess Peak advances.

The absence of any moraines in the cirques between Matthes and Recess Peak deposits also demonstrates that the Matthes glaciers of the last millennium are the most extensive of the Holocene. Nevertheless, detailed relative weathering measurements on moraines and rock glaciers mapped as Matthes-age suggest that some, in fact, may have formed before the last millennium (Yount *et al.*, 1979; 1982). However, these "older Matthes" moraines are not physically separate from nor significantly or consistently larger than those of the last thousand years, being preserved only locally where subsequent, more extensive Matthes advances have fortuitously not buried them. A recent Holocene climate record based on stable isotopes in tree rings from the White Mountains, immediately east of the southern Sierra Nevada, indicates that the last 500 years were the coolest of the past 8000 years (Feng and Epstein, 1994), supporting the contention that the Matthes advance was the largest widespread advance of the Holocene. Tree-line fluctuations in the southern Sierra Nevada (Scuderi, 1987a) and the White Mountains (LaMarche, 1973) are also consistent with this inference.

Matthes debris-covered glaciers may preserve an integrated late-Holocene history of these Little Ice Age advances. Many of Yount *et al.*'s (1982) "older Holocene" RW sites occur on rock glaciers that are still actively advancing, and thus may be more properly thought of as "modern," though potentially long-lived, glaciers. Indeed, many debris-covered Matthes glaciers are still advancing despite significant shrinkage or disappearance of their accumulation zones (Clark *et al.*, 1994a, b).

Early direct observations of Sierran glaciers indicate that in the late 19th and early 20th centuries, glaciers were at or very close to the maximum extent of their Matthes moraines (LeConte, 1873; Muir, 1873; Russell, 1885; Gilbert, 1904; Matthes, 1930; Matthes, 1936; Matthes, 1939). Whiting (1985) cited moraine morphology and lichenometry as evidence that the small, now-extinct Matthes glaciers in Desolation Wilderness (Fig. 3.7) were probably active until this century. Together, these observations indicate that the snowline depression of the last century in the Sierra Nevada was at or near the maximum for the entire Holocene interglaciation.

Regional ELA depressions during Recess Peak and Matthes advances in the study area show that both events involved significant change of climate relative to modern conditions. Modern and Matthes-age ELAs estimated by Gillespie (1991) in Sierran cirques that contained both deposits indicates that Matthes ELAs were typically ~60 m lower than modern ELAs. My work shows that Recess Peak ELAs average ~150 m lower than Matthes ELAs in the same cirques, or roughly 210 m below the ELAs of the modern glaciers. Although there is considerable scatter about these averages, and the ELAs of the modern, rapidly retreating glaciers do not represent steady-state conditions, the numbers indicate that the Recess Peak advance resulted from a lowering of regional snowline more than twice that which occurred during the height of the Little Ice Age. ELAs in the same drainages during the Tioga stade were generally 700-800 m below the modern ELAs (Gillespie, 1991; Burbank, 1991). Burbank estimated that the ELAs of the small, protected modern Sierran glaciers were about 100-200 m below the regional, or climatic snowline, and therefore total snowline lowering during maximum glaciation was closer to 900-1000 m relative to modern conditions. Thus, the Recess Peak advance resulted from snowline-lowering of ~20% that of full-glacial conditions. Snowline depression during the Little Ice Age was generally less than 10%. The large local influence of enhanced shading and accumulation in these small glaciers preclude quantitative estimates of the magnitude of precipitation and temperature changes from modern values that are represented by these shifts in snowline (e.g., Leonard, 1989).

The relatively consistent difference between Matthes and Recess Peak ELAs along the Sierra Nevada crest suggests that the regional climates that caused each advance were similar in character, and did not have markedly different precipitation and temperature patterns. The consistent spatial distribution of, and relation between, the two sets of deposits along the crest of the range also indicate that the local age control on the moraines presented here is representative of deposits throughout the range. Recess Peak deposits along the range crest therefore likely represent a single event that affected the entire range rather than being a lumping of several distinct events.

Whereas the age of the maximum Matthes advance has been restricted to the last 700 years (Wood, 1977; Yount *et al.*, 1982), the timing of the Recess Peak advance has remained controversial. Although it has generally been regarded as a late-Neoglacial event (e.g., Birman, 1964; Curry, 1969; Scuderi, 1990, p. 84), the new basal dates from cores of alpine lakes behind Recess Peak recessional moraines demonstrate that the advance was of late-Pleistocene age. The oldest basal dates, 11,190 ± 70 ^{14}C yr B.P. (13,115-13,180 cal yr B.P.) and 10,880 ± 60 ^{14}C yr B.P. (12,730-12,880 cal yr B.P.) for gyttja from cores BL-2 and BL-1, respectively, are minimum ages and indicate that the Recess Peak advance had *ended* before the *onset* of the European Younger Dryas climatic reversal (~11,000 ^{14}C yr B.P., Mangerud *et al.*, 1974; 12,700-12,800 ± 100-250 sidereal years, Johnsen *et al.*, 1992; Alley *et al.*, 1993)(Fig. 3.12).

Although there are uncertainties in the dating of the onset of the Younger Dryas, which may allow a small statistical overlap with the minimum date for the end of the Recess Peak advance, several lines of evidence indicate that the events are of different ages: (1) the oldest dates come from inside the innermost recessional Recess Peak moraines, and thus postdate the last significant stillstand or readvance (rather than the maximum extent) of the glacier before it disappeared; (2) the sediments that provide the oldest dates occur above the initial organic (e.g., postglacial) sediments within the cores,

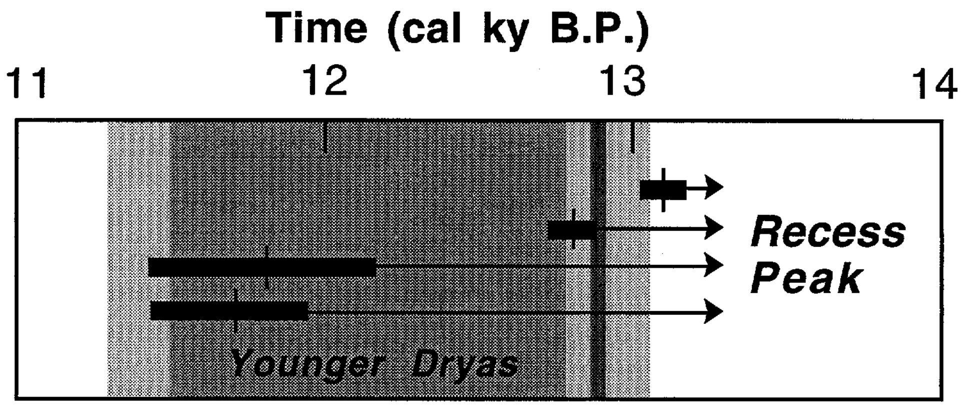


Figure 3.12: Time-line showing relationship between the four oldest limiting radiocarbon dates (calibrated according to Stuiver and Reimer, 1993) of Recess Peak advance and sidereal age of Younger Dryas cooling recorded in Greenland ice cores (medium gray band; Johnsen *et al.*, 1992; Alley *et al.*, 1993). Black boxes indicate 1σ age range of AMS dates; arrows indicate that all ages are minima for end of Recess Peak advance. Light gray bands show outer limits of combined error of Johnsen *et al.* (1992) and Alley *et al.* (1993). Dark gray line indicates calibrated radiocarbon age estimate of onset of Younger Dryas cooling in Scandinavia (Mangerud *et al.*, 1974). Minimum age for end of Recess Peak advance and maximum age of onset of Greenland Younger Dryas event overlap only at outer limits of both error fields.

and therefore may significantly postdate the end of the Recess Peak advance (e.g., Davis and Davis, 1980); at the least, they represent some period of time *after* the retreat of the glacier; (3) core BL-2, from which the oldest date came, is not from the deepest part of the lake; (4) the many AMS-radiocarbon dates of gyttja, peat, and macrofossils I have obtained from similar high-altitude lakes in the Sierra Nevada are all internally consistent (e.g., Fig. 3.8, Table 3.1; D. Clark, unpublished data), indicating that the ^{14}C ages in the cores are therefore accurate; (5) with the one exception in core BL-2 explained above, dates of adjacent macrofossils and peat and adjacent macrofossils and gyttja are indistinguishable within the 1σ analytic error (typically <1%), also supporting the accuracy of the ^{14}C ages in the cores; (6) the silicic bedrock in the basins upstream of the coring sites, high in the accumulation zones of the Wisconsin glaciers, must have been thoroughly stripped of any organic material by glacial plucking during the thousands of years of Tioga glacierization prior to the Recess Peak advance; thus there would have been no significant sources of "old" carbon to contaminate the lakes and sediments I cored, and all dates from the cores are minima for the Recess Peak advance.

Together, these observations indicate that small Sierran lakes are stable and uncontaminated reservoirs for carbon-bearing sediments, and support the validity of the oldest dates. Furthermore, although some of the large Scandinavian glaciers and pollen profiles on which the Younger Dryas event was originally defined (e.g., Mangerud *et al.*, 1974) may have responded more slowly to a change in climate than Sierran cirque glaciers; changes in accumulation and isotopes recorded in the Greenland ice cores are nearly instantaneous and can be resolved to annual to decadal levels during the late-glacial period (e.g., Johnsen *et al.*, 1992; Alley *et al.*, 1993). Therefore, to have the Recess Peak advance and the Younger Dryas cooling actually responding to the same climatic forcing would require 1) that the oldest date from the Baboon Lakes cores is too old by nearly 2000 yr, or that the dates for the onset of the Younger Dryas cooling are too young, both for unspecified reasons, and 2) the Recess Peak was a nearly instantaneous event that began and ended well before the height of the Younger Dryas cooling in the North Atlantic. Such a scenario is unreasonable.

The new limiting dates for the Recess Peak advance, in conjunction with the consistent absence of moraines between those of the Recess Peak and Matthes advances, therefore imply that any glaciers which may have existed in the Sierra Nevada during the Younger Dryas interval, or subsequently, were smaller than those of the Matthes advance of the last millennium. This finding requires a reevaluation of the classic sequence of late-glacial and Holocene glaciation for the range (Fig. 3.4). Moreover, it contradicts a recent correlation of moraines in the eastern Sierra Nevada with the Younger Dryas on the basis of ^{36}Cl surface-exposure ages of morainal boulders (Phillips *et al.*, 1993). Dates that bracket a distinct light-greenish silt near the base of core BL-1 (Fig. 3.8D) indicate that this sediment accumulated between ~10,900 and 10,100 ^{14}C yr B.P. (12,800-11,700 cal yr B.P.), closely coincident with the Younger Dryas chronozone. The distinguishing color, grain-size, and organic content of the silt layer relative to the sediments above and below suggest that it represents a distinct environmental, and thus possibly climatic, event during its deposition. The sharp upper boundary of the unit suggests that the event ended rapidly. Palynologic work (in progress) on the cores may help refine the climatic interpretation of this interval. Possibly the silt records a climatic oscillation of Younger Dryas age.

IMPLICATIONS

The new age limits on Sierran deglaciation presented here have several regional implications. In particular, they raise questions about the extent and expression of Younger Dryas cooling in the Northern Hemisphere. Several recent studies concluded that late-glacial advances in several other mountain ranges in western North America may have coincided with the Younger Dryas cooling in the North Atlantic. For example, Evenson *et al.*, (1994) and Gosse *et al.* (1995a) use measurements of *in situ* cosmogenic ^{10}Be and ^{26}Al in morainal boulders to claim that the Titcomb Lakes moraines in the Wind River Range, Wyoming, correlate with the Younger Dryas interval. As discussed in detail in the next chapter, the new data on Sierran deglaciation substantially changes the late-glacial production rates of, and thus exposure ages calculated from, these two cosmogenic isotopes. As a result, it is possible that the Titcomb Lakes moraines are substantially older than reported by Gosse *et al.* (1995a) and actually predate the Younger Dryas event. Such a conclusion is consistent with Zielinski and Davis (1987), who obtained radiocarbon dates of 11,700 ± 710 (12,970-14630 cal yr B.P.) and 11,400 ± 630 ^{14}C yr B.P. (12,700-14,040 cal yr B.P.) for lacustrine sediments associated with the Temple Lake moraines in the Wind River Range, which correlate with those dated by Gosse *et al.* (Hall, 1989). The older dated sediments overly outwash from the outer Temple Lake moraine, whereas the younger dated sediments overly outwash inside the inner moraine. Thus, the dates appear to give minimum ages for the maximum position and final retreat of the Temple Lake glacier, respectively. In particular, the older date, although refered to by Fall *et al.* (1995, p. 399) as a "maximum" age for the outer moraine, is stratigraphically a close minimum limiting age for the formation of the moraine.

Although both radiocarbon dates are conventional and measured on bulk sediments, Zielinski and Davis, and more recently Fall *et al.* (1995), state that there are few sources for contamination (particularly from "old" carbon), and that the residence time of water in the lakes (i.e., the carbon reservoir effect) is probably no more than a few hundred years. Any reservoir effects, particularly for the lakes formed inside the moraines, would be less when the glacier first retreated, because they were newly formed and water through-put would presumably be high during deglaciation. The absence of sources of "old" carbon, coupled with the large size of the dated sediment samples (integrating 30 and 9 cm of core length, respectively; Zielinski and Davis, 1987), indicate that the dates are probably accurate minima for the advance and retreat of the glacier. Therefore, the age of the Temple Lake advance, as dated by Zielinski and Davis, is consistent with a pre-Younger Dryas glacial event, possibly equivalent to the Recess Peak advance in the Sierra Nevada.

Further supporting this conclusion, J. Heine (1995, personal communication) finds evidence for advances that predate and possibly postdate the Younger Dryas interval in the Cascade Mountains near Mt. Rainier, Washington. Basal radiocarbon ages on sediment cores from bogs dammed behind late-glacial moraines indicate that the older of a regional doublet of moraines formed before ~11,100 ^{14}C yr B.P. (~13,000 cal yr B.P.). Minimum limiting ages for the younger moraine of the doublet indicate it formed before ~9000 ^{14}C yr B.P. (~10,000 cal yr B.P.)(Heine, 1995, personal communication). Although Heine believes sediment relationships in the cores indicate that the latter date is a close minimum, it is possible that the younger moraine set is Younger Dryas age.

Dated sediments in Crowfoot Lake in the Canadian Rockies indicate that the type Crowfoot moraines likely did form during Younger Dryas time (Reasoner *et al.*, 1994; Osborn and Gerloff, *in press*). An increase in cold-water foraminifera in the waters off the coast of British Columbia also occurred at this time (Mathewes *et al.*, 1993), lending credence to the Crowfoot-Younger Dryas correlation. Taken at face value, these results suggest a regionally variable climate in North America during Younger Dryas time, with glaciers advancing in Canada, but not in the Sierra Nevada. Unfortunately, adequate glacial or other climatic data for this period are absent in the intervening region (i.e., Cascades, Great Basin, American Rockies), leaving the nature and cause of this possible difference in climatic response unresolved. However, the regional difference does imply that the response to the Younger Dryas climatic event was not everywhere the same, or that the effect was not global. Certainly it calls into question recent attempts to connect late-Pleistocene moraines in North America to Heinrich events and the Younger Dryas (e.g., Easterbrook, 1994; Davis, 1994; Clark and Bartlein, 1995). Such results also suggest that records of local glacier advances outside the North Atlantic that apparently coincide with the Younger Dryas (e.g., Denton and Hendy, 1994; Rodbell *et al.*, 1994) do not necessarily signify global climatic forcing for the event recorded in the North Atlantic.

CONCLUSIONS TO CHAPTER 3

The late-Pleistocene and Holocene glacial record in the Sierra Nevada is well preserved, yet the timing and climatic significance of the deposits has been unclear. The mapping presented here confirms that two cirque-glacier advances originally defined by Birman (1964) in the central Sierra Nevada (the Matthes and the Recess Peak) occur throughout the range from the southern limit of glaciation as least as far north as Lake Tahoe. Other work in progress indicates that they are also the only two regional advances since stagnation of the much larger late-Wisconsin Tioga glaciers. The spatial, stratigraphic, and morphologic relation between the two advances remain consistent wherever both occur in the range, supporting my correlations. Reconstructions of the glaciers along the range crest indicate that during the Matthes and Recess Peak advances, ELAs fell by ~10% and ~20%, respectively, of the maximum late-Pleistocene ELA change throughout the range. The relatively consistent difference between the snowline lowering of the two advances suggests that the climatic events that caused each were similar in character and differed mainly in magnitude.

Dated sediment cores from small tarns formed behind Recess Peak moraines demonstrate that it was a late-Pleistocene event that ended before ~13,100 cal yr B.P., not a late-Holocene event as previously thought. These minimum dates further demonstrate that the advance also predated the North Atlantic Younger Dryas cooling event. Late-Holocene Matthes glaciers reached their maximum extents within the last 600-700 cal yr B.P., during the Little Ice Age. Earlier Neoglacial advances, if any, apparently did not extend further than did the Matthes glaciers. Thus, there is no evidence for glaciation greater than the Little Ice Age maximum during the ~12,000 years that separate the Recess Peak and Matthes advances.

The apparent absence of a glacier advance in the Sierra Nevada during Younger Dryas time implies that the Sierra Nevada was too warm and/or too dry to support even small glaciers equivalent to Matthes in size. However, evidence for a Younger Dryas advance in the Canadian Rockies suggests regionally variable responses of local climate in North America during this event. The Sierran record also raises questions about the global extent of the Younger Dryas cooling, at least as a glacial event.

The new age limit on the Recess Peak advance is a minimum for deglaciation of the Sierra Nevada. Dates from sediment cores in other lakes in the range support the Recess Peak age limits, and suggest that the range may have been deglaciated to Recess Peak ice limits by 14,000-15,000 cal yr B.P. (Clark *et al.*, 1995 (in press)). These new constraints on the timing of deglaciation of the Sierra Nevada may require revision of current production rates for cosmogenic ^{10}Be and ^{26}Al during the late-glacial and Holocene periods. These rates were established from glacier-polished bedrock in the range that lies outside Recess Peak ice limits and which was assumed to have been deglaciated 11,000 cal years ago. Such a change in production rates casts doubt on the accuracy of ^{10}Be and ^{26}Al exposure ages obtained using current production rates, including those used to correlate glacial events with short-lived climatic events. Thus moraines in the Wind River Range previously correlated with the Younger Dryas event on the basis of ^{10}Be exposure ages may actually predate the event; increasing these ages by ~20%, as suggested by the new Sierran chronology, brings these dates into agreement with limiting radiocarbon ages from nearby lake cores, and implies that the moraines may correlate with the pre-Younger Dryas Recess Peak advance.

CHAPTER 4: IMPLICATIONS OF REVISED SIERRAN DEGLACIAL CHRONOLOGY FOR SURFACE EXPOSURE DATING

INTRODUCTION TO CHAPTER 4

Measurement of nuclides produced by cosmic-ray bombardment in surficial deposits and exposed bedrock is rapidly becoming an important tool in paleoclimatology, geomorphology, and other physical sciences. In particular, such measurements have been used to determine the age of otherwise numerically undateable landforms, to estimate rates of surface processes, and to correlate geologic, climatologic, and tectonic events (e.g., Bierman, 1994; Cerling and Craig, 1994a; Nishiizumi *et al.*, 1993). As the precision of isotopic measurements, and thus the apparent precision of calculated exposure ages, has improved in the past several years, studies have increasingly attempted to date and correlate short-lived, particularly climate-related, events across wide regions (e.g., Clark and Bartlein, 1995; Easterbrook, 1994; Evenson *et al.*, 1994; Gosse *et al.*, 1995a, 1995b; Phillips *et al.*, 1990; Phillips and Zreda, 1992). The reliability of such dating, and of any resulting correlations, depend directly upon the accuracy and precision of the isotope production rates from which exposure ages are calculated.

Typically, time-integrated, "average" isotope production rates are determined empirically from geomorphic surfaces of "known" age (Cerling and Craig, 1994b). These production rates are assumed to be constant when calculating exposure ages. Because geomagnetic field strength, and therefore isotope production rates at latitudes <50° have varied in the past (Cerling and Craig, 1994a, b; Kurz *et al.*, 1990; Mazaud *et al.*, 1991; McElhinny and Senanayake, 1982, Weeks *et al.*, 1995), the use of average production rates will, in most cases, generate systematic errors in calculated cosmogenic exposure ages. Furthermore, geomagnetically induced production rate variability and the associated systematic errors in calculated exposure ages are complexly related to the latitude, altitude, and exposure age of both sample and calibration sites.

My research in the Sierra Nevada, and that of my colleagues elsewhere (Clark *et al.*, 1995 (in press)), show that uncertainties in ages of calibration surfaces and geomagnetic field strength variations contribute substantial and typically unstated errors to cosmogenic age estimates. New radiocarbon ages demonstrate that the assumed exposure age of glaciated Sierra Nevada rocks used to calibrate late-Pleistocene and Holocene production rates for ^{10}Be and ^{26}Al (Nishiizumi *et al.*, 1989) is probably at least 20% too young. Preliminary ^{10}Be data from the Laurentide terminal moraine in New Jersey support this finding and along with the Sierran radiocarbon data imply that many exposure ages calculated using previously accepted production rates are also too young. Calculations based on proxy records of geomagnetic field strength (Clark, 1995 (in press)) suggest that even if calibration sites were correctly dated, field-strength and attendant production-rate variations may generate age errors >20% over the past ~100,000 yr for all *in situ* isotope systems if time-averaged production rates are used to calculate exposure ages.

These findings indicate that inaccuracy in calculated cosmogenic exposure ages is significantly greater than previously suggested by analytic precision (typically 3-8%) or by the lowest variance so far associated with multiple samples from a monogenetic landform (4%, Gosse *et al.*, 1995b, p. 1330). The variety and magnitude of systematic uncertainties, illustrated by the Sierran data and the calculations in Clark *et al.* (1995 (in press)), currently prohibit rigorous cosmogenic dating or correlation of short-lived events such as the Younger Dryas climatic reversal or Heinrich events.

^{10}BE AND ^{26}AL PRODUCTION RATES AND A REVISED GLACIAL CHRONOLOGY FOR THE SIERRA NEVADA

In situ isotope production rates have been estimated in two ways: theoretically using excitation functions (Lal, 1991), and empirically by analyzing samples exposed to the cosmic ray flux for discrete time periods. For the latter method, Yokoyama *et al.* (1977) estimated a variety of short-term (years) production rates on the basis of Na-isotopes produced in Al-targets at high altitudes. Long-term production rates (millennia), which I discuss in this chapter, have typically been calibrated on glacially polished bedrock, morainal boulders, or lava flows. Several assumptions underlie such long-term calibrations: 1) the calibration site was instantaneously exposed at a known time in the past; 2) the site has no isotopic inheritance from previous exposure; 3) the site has not been significantly buried after exposure (e.g., by loess, till, or snow); and 4) the sampled surface has not eroded since exposure. In this chapter, I discuss the effect that changes in the first criterion may have on previously accepted production rates for ^{10}Be and ^{26}Al.

Nishiizumi *et al.* (1989) made the first calibration of *in situ* cosmogenic ^{10}Be and ^{26}Al by collecting samples of glacially polished granite in the Sierra Nevada and by assuming that their three sample sites (37°59' N, 3180 m altitude; 37°25' N, ~3550 m; 38°52' N, ~2440 m; Fig. 4.1) were deglaciated 11,000 calibrated (or sidereal) years ago (11,000 cal yr B.P.). This exposure-age estimate was based on minimum-limiting conventional radiocarbon ages of ~10,000 ^{14}C yr B.P. for basal bulk sediments from two sites, a lake and a meadow, dammed behind Tioga recessional moraines (Table 4.1; Adam, 1967; Mezger and Burbank, 1986). On the basis of this estimated age, Nishiizumi *et al.* reported sea-level, high-latitude production rates in quartz of 6.03 and 36.8 atoms g^{-1} $year^{-1}$ for ^{10}Be and ^{26}Al, respectively, assuming that the modern geomagnetic latitude of the samples (44°) was representative of their geomagnetic latitude since exposure. Although Nishiizumi *et al.* (1989, p. 17,913) reported uncertainties in both the timing of

Figure 4.1. Map of Sierra Nevada showing approximate Tioga (late-Wisconsin) glacier limits, and sample sites discussed in text. Red diamonds: coring sites presented in this paper; blue triangles: coring sites used by Nishiizumi *et al.* (1989); yellow squares: ^{26}Al- and ^{10}Be-calibration sample sites of Nishiizumi *et al.* Glaciated region includes nunataks. Sierra Nevada crest shown by dashed gray line. Modified from Wahrhaftig and Birman (1965).

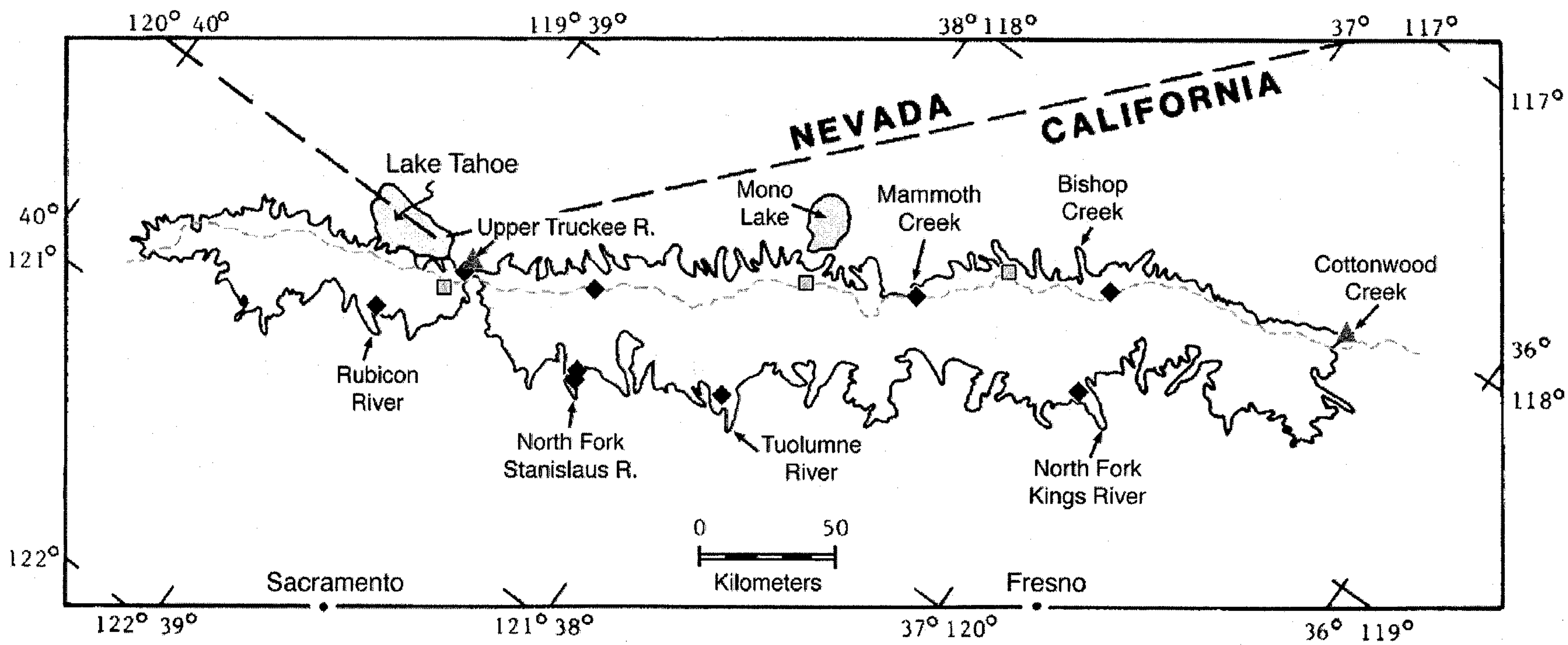

120° 40°
119° 39°
38° 118°
37°
117°
117°
NEVADA
CALIFORNIA
Lake Tahoe
Upper Truckee R.
Mono
Lake
Mammoth
Creek
Bishop
Creek
Cottonwood
Creek
40°
121°
36°
118°
Rubicon
River
North Fork
Stanislaus R.
Tuolumne
River
North Fork
Kings River
0
50
Kilometers
122°
Sacramento
Fresno
122° 39°
121° 38°
37° 120°
36° 119°

Table 4.1. Minimum Limiting Radiocarbon Dates on Deglaciation of the Sierra Nevada, California

Location	N Latitude	W Longitude	Altitude (m)	Geologic Setting of Cored Lake	Dated Material	Basal Date (^{14}C yr B.P.)	Basal Date (cal yr B.P., 1-σ)[13]
Swamp Lake, N. Fork Stanislaus River[1]	38°22'	120°08'	1957	Dammed within Tioga lateral moraines	outwash silts	15,565 ± 820†	**17,638-19,294**
Lake Moran, N Fork Stanislaus River[2]	38°30'	120°08'	2017	Dammed within Tioga lateral moraines	organic silt and sand	14,750 ± 500†	**17,080-18,180**
Swamp Lake, Tuolumne River[3]	37°57'	119°49'	1554	Dammed within Tioga recessional moraines	organic sediments	13,690 ± 340†	**15,948-16,838**
"Granite Gorge" pond, N. Fork Kings River[4]	36°59'	119°00'	1980	Tarn upstream of Tioga recessional moraines	organic silt (~550 cm sediment depth)	15,230 ± 470†	**17,652-18,604**
"Granite Gorge" pond, N. Fork Kings River[5]	36°59'	119°00'	1980	Tarn upstream of Tioga recessional moraines	gyttja (267 cm sed. depth)	11,700 ± 60‡	**13,531-13,766**
					gyttja (259 cm sed. depth)	11,690 ± 60‡	**13,520-13,754**
Bunker Lake, Rubicon River[6]	39°03'	120°23'	1995	Dammed by Tioga recessional moraines	gyttja	11180 ± 180†	**12,910-13,280**
Upper Echo Lake, Upper Truckee River[7]	38°50'	120°03'	2260	Tarn upstream of Tioga recessional moraines	organic silt	11,100 ± 70‡	**12,929-13,094**
Highland Lks, N Fork Stanislaus River[8]	38°29'	119°48'	2625	Tarn in Tioga accumulation zone	organic silt	13,270 ± 200†	**15,528-16,139**

(continued on next page)

Table 4.1 (continued).

Location	N Latitude	W Longitude	Altitude (m)	Geologic Setting of Cored Lake	Dated Material	Basal Date (^{14}C yr B.P.)	Basal Date (cal yr B.P., 1-σ)[13]
Barret Lake, Mammoth Creek[9]	37°36'	119°00'	2816	Dammed by Tioga recessional moraines	organic silt	11,730 ± 430†	**13,204-14,204**
Baboon Lk. 1, Middle Fork Bishop Creek[10]	37°10'	118°37'	3385	Cirque tarn upstream of Recess Peak moraines	gyttja	11,190 ± 70‡	**13,015-13,187**
Baboon Lk. 2, Middle Fork Bishop Creek[5]	37°10'	118°37'	3385	Cirque tarn upstream of Recess Peak moraines	gyttja	10,880 ± 60‡	**12,727-12,882**
Osgood Swamp∞, Upper Truckee River[11]	38°51'	120°02'	1985	Bog dammed by Tioga recessional moraine	peat (30-40 cm above outwash)	9990 ± 800†	**10,050-12,721**
Cottonwood Lks∞, Cottonwood Creek[12]	36°28'	118°10'	2880	Meadow behind Tioga recessional moraine	peat	10,640 ± 160†	**12,383-12,738**

∞ = sites used for calibration of Nishiizumi *et al.* (1989); † = Conventional radiocarbon age; ‡ = AMS radiocarbon age

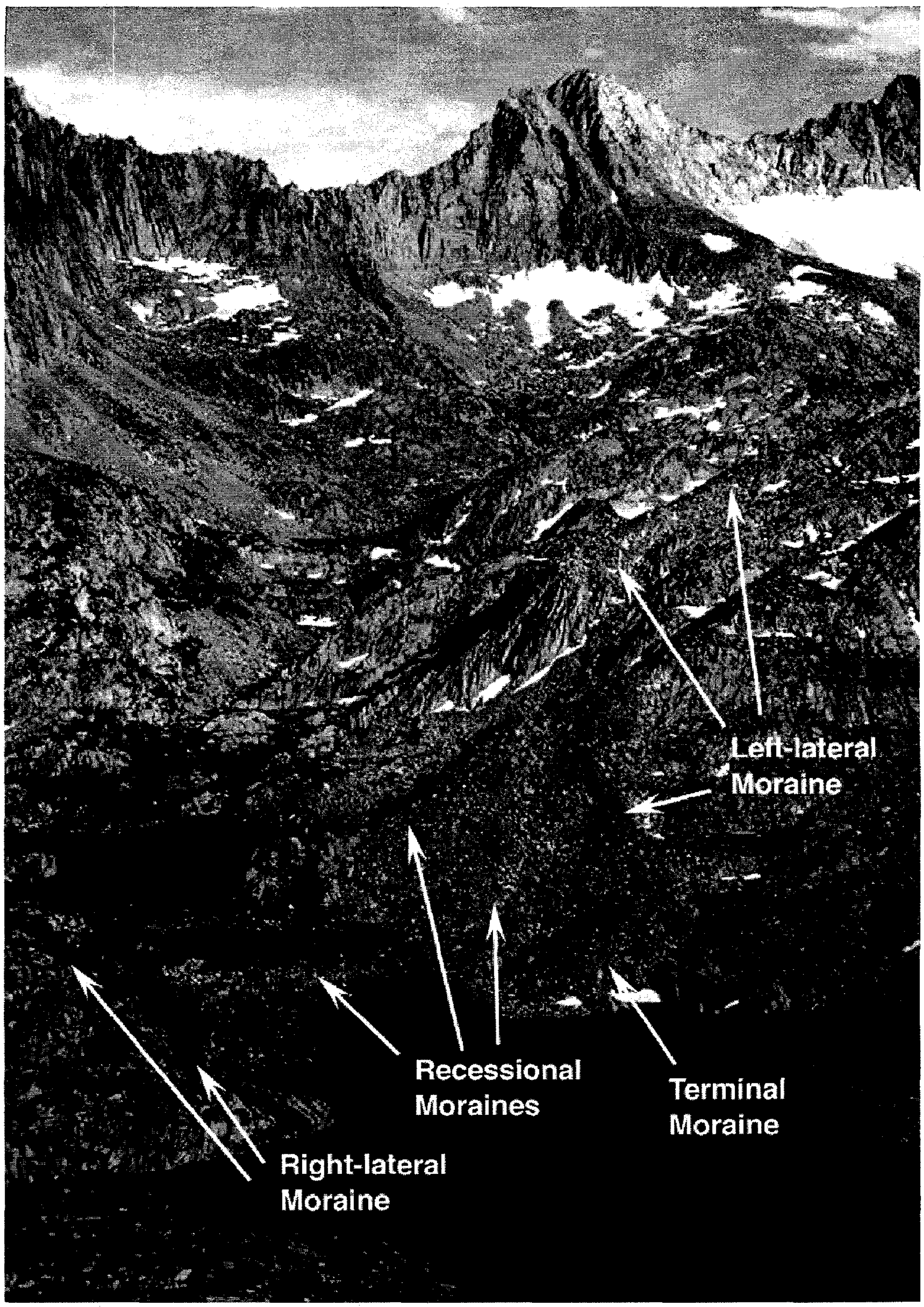
Sources Table 4.1: 1. Batchelder (1980); 2. Edlund and Byrne (1991); 3. Smith and Anderson (1992); 4. E. Edlund (1994, written communication); 5. Clark, this study; 6. Edlund (1993); 7. Adam (1985); 8. Byrne (1994; written communication); 9. Anderson (1990); 10. Clark and Gillespie (in press); 11. Adam (1967); 12. Mezger and Burbank (1986); 13. Stuiver and Reimer (1993), Bard *et al.* (1990).

deglaciation (± 5%) and the temporal stability of the geomagnetic field (potential production rate decrease of ~15%), most subsequent ^{10}Be and ^{26}Al studies rely on Nishiizumi *et al.'s* production rates without acknowledging the full geologic and geomagnetic uncertainties inherent in the original calibration, despite commonly discussing in detail other sources of potential error (e.g., Evenson *et al.*, 1994; Gosse *et al.*, 1993; Macchiaroli *et al.*, 1994; McCuaig *et al.*, 1994; Nishiizumi *et al.*, 1993).

New ^{14}C ages for Sierra Nevada deglaciation suggest that the production rates of Nishiizumi *et al.* (1989) are at least 20% too high because the age of exposure they chose was too young. Calibrated ages (Bard *et al.*, 1990; Stuiver and Reimer, 1993) of twelve radiocarbon dates for basal or near-basal lake sediments from cores of ten postglacial lakes show that Sierra Nevada deglaciation was underway by 16,000-19,000 cal yr B.P. and that the range was largely deglaciated before ~13,100 cal yr B.P. (Table 4.1). All coring sites listed in Table 4.1 lie inside the maximum Tioga ice limits, as do the two used by Nishiizumi *et al.*, and thus provide minimum ages for onset of deglaciation (Fig. 4.1). It is the highest sites, however, near the crest of the range, that provide minimum ages for the complete deglaciation of the range and thus minima for the exposure age of sites sampled by Nishiizumi *et al.* for calibration. The new coring sites listed in Table 4.1 encompass the region sampled by Nishiizumi *et al.*; furthermore, many are closer to the ^{10}Be and ^{26}Al calibration sites than are the two coring sites cited by Nishiizumi *et al.* Together, the twelve radiocarbon dates indicate that Tioga deglaciation in the Sierra Nevada occurred at least 2000 cal yr before 11,000 cal yr B.P.

The timing of full deglaciation is particularly important for evaluating the actual exposure age of Nishiizumi *et al.*'s calibration sites and is demonstrated by a pair of high-precision AMS radiocarbon dates from Baboon Lakes, a series of tarns formed behind a set of small moraines in the headwaters of Middle Fork Bishop Creek (Fig. 4.2). The

Figure 4.2. Recess Peak moraines at Baboon Lakes coring site. Glacier flowed from cirques in background and out of picture to right into lake in foreground. Left-lateral moraine dives into lake in center foreground; right-lateral moraine covers crest of rounded, tree-covered bedrock ridge to left of upper lakes, out of photo. Recess Peak recessional moraines occur only between outer moraine and uppermost visible lake. Basal radiocarbon dates from two small tarns inside innermost recessional moraines, to left of upper lake just out of photo, provide minimum limiting age for the end of the Recess Peak glaciation, and thereby total deglaciation of the range, of 11,190 ± 70 ^{14}C yr B.P. (~13,100 cal yr B.P.). This age is also a minimum for deglaciation of ^{26}Al and ^{10}Be calibration sample sites of Nishiizumi *et al.* (1989), which lie downstream of similar moraines in other drainages.


Left-lateral
Moraine
Recessional
Moraines
Terminal
Moraine
Right-lateral
Moraine

moraines surrounding these lakes correlate with the Recess Peak advance (Birman, 1964; Clark *et al.*, 1994; Clark and Gillespie, 1994, 1995 (in press)).

Clark and Gillespie (1994, 1995 (in press)) have mapped Recess Peak deposits along the crest of the Sierra Nevada from Lake Tahoe to the southern limit of glaciation (Fig. 4.1), and the sites sampled by Nishiizumi *et al.* (1989) all lie downstream of Recess Peak moraines in those drainages. Thus all sites used to calibrate the production of *in situ* ^{10}Be and ^{26}Al were deglaciated *before* the onset of the Recess Peak advance. The minimum limiting dates from Baboon Lakes, in conjunction with the other dates in Table 4.1, indicate that Tioga ice in the Sierra Nevada retreated to the crest of the range, and then advanced and disappeared during Recess Peak time, all before ~13,100 cal yr B.P. (Clark, 1976; Clark and Gillespie, 1994, 1995 (in press)).

The accuracy and precision of the Baboon Lakes dates are supported by seven other internally consistent AMS radiocarbon dates higher in the two cores (Chapter 3). Moreover, dates of adjacent gyttja, wood chips, and charcoal are indistinguishable within analytic error (typically <1%, 1σ), indicating the gyttja has been neither disturbed nor contaminated since deposition. Sixteen internally consistent dates on macrofossils, gyttja, peat, and tephra from cores of 4 other similar tarns elsewhere in the Sierra Nevada further support the accuracy of such lake sediments for radiocarbon dating (Clark and Gillespie, 1994, 1995 (in press)). The lack of any significant sources of "old" carbon in the granitic basins upstream of the lakes indicates that all dates of the Recess Peak cores are minima for deglaciation.

Considering the new age limits for retreat of Recess Peak glaciers and the other minimum limiting dates for substantial Tioga deglaciation (e.g., Highland Lakes near the Sierran crest deglaciated by 15,500 cal yr B.P.; R. Byrne, written communication, 1994; Table 4.1), I infer that late-Wisconsin deglaciation of the Sierra Nevada was completed by 13,000 to 15,000 cal yr B.P., rather than 11,000 cal yr B.P. Therefore, exposure ages calculated from Nishiizumi *et al.*'s (1989) production rates are systematically too young unless compensated by other unrecognized errors. Using the conservative estimate that Nishiizumi *et al.*'s sites were deglaciated 13,000 cal yr B.P., integrated production rates (sea level, high latitude) for ^{10}Be and ^{26}Al would be 5.10 and 31.1 atoms g^{-1} year^{-1}, respectively. Considering that Nishiizumi *et al.*'s sites were probably exposed earlier than the minimum age for the Recess Peak advance (Clark and Gillespie, 1994; in press), I feel 14,000 cal yr B.P. is a better estimate for their deglaciation, indicating production rates of 4.74 and 28.9 atoms g^{-1} year^{-1} for ^{10}Be and ^{26}Al, respectively.

INDEPENDENT SUPPORT FOR REVISED PRODUCTION-RATE CALIBRATION

Two other studies lend significant independent support for the revised chronology of Sierran deglaciation: one that estimated production rates for ^{10}Be in rocks scoured by the late-Wisconsin Laurentide ice sheet; and one that calculates the effects on in-situ production rates of past fluctuations in the strength of the Earth's geomagnetic field. Both studies support my inference above that current production rates for ^{10}Be and ^{26}Al are too high by ~20%.

Larsen and Bierman (in Clark *et al.*, 1995 (in press); Larsen *et al.*, 1995) measured ^{10}Be concentrations in a suite of rock samples from on and near the Late-Wisconsin terminal moraine in west-central New Jersey. Radiocarbon dates, stratigraphic evidence, and palynological studies from lacustrine and bog-bottom sediments near the moraines provide minimum limiting ages for deglaciation of the area, and indicate that the analyzed rocks were exposed ~21,500 cal yr B.P.

They analyzed 15 rock samples from the Laurentide moraine, including five replicates, at Lawrence Livermore National Laboratory. The average concentrations of ^{10}Be in the five boulder and five bedrock samples are similar, and average $1.02 \pm 0.10 \times 10^5$ atoms g^{-1} (1σ). Replicate measurements of five samples agree within 1σ

uncertainties. If 2σ errors are considered, all ten bedrock and boulder samples are indistinguishable. By taking the exposure age of these samples to be 21,500 cal yr (as indicated by the bog dates), they calculate a preliminary sea-level, high-latitude production rate of 4.76 ± 0.47 ^{10}Be atoms g^{-1} yr^{-1} (1σ of mean). This provisional production rate agrees closely with the revised Sierran calibration.

The accuracy and precision of calculated production rates are affected not only by the assumed exposure age of calibration sites as shown above, but also by temporal changes in the strength of Earth's geomagnetic field, which modulates the flux of incoming primary cosmic radiation; e.g., high field strength diminishes the abundance of cosmic rays penetrating the atmosphere and reaching Earth's surface and thus lowers *in situ* isotope production rates. Many studies have documented field strength changes during the Pleistocene and the impact of these changes on isotope production rates, primarily in the atmosphere (Bard *et al.*, 1990; Mazaud *et al.*, 1991; McElhinny and Senanayake, 1982; Meynadier *et al.*, 1992; Weeks *et al.*, 1995). Only a few studies have explicitly considered the effect of field strength changes on the production of cosmogenic nuclides *in situ* (Kurz *et al.*, 1990; Cerling and Craig, 1994a, b).

In order to estimate the magnitude of geomagnetically induced exposure age errors and to demonstrate that model corrections can be applied to reduce these errors, Bierman (in Clark *et al.*, 1995 (in press)) calculated the effect of changing geomagnetic field strength on *in situ* production rates of cosmogenic isotopes using several simple models available in the literature. The resulting calculations (described in Clark *et al.*, 1995 (in press)) indicate that, for the current Sierran calibration (Nishiizumi *et al.*, 1989), an age-error of ~10% occurs for exposure ages greater than 40,000 yr.

Although the accuracy of paleointensity records on which our calculations (Clark *et al.*, 1995 (in press)) are based is uncertain (Raisbeck *et al.*, 1994), many late-Pleistocene records indicate that substantial changes in field strength have occurred, that to the first-order the records from widely separated areas are consistent in the sign and magnitude of those changes (e.g., Weeks *et al.*, 1995), and therefore that a static late-Quaternary paleomagnetic field is unlikely. Even if the magnitude and timing of field strength changes are later shown to differ from the paleointensity records Clark *et al.* (1995 (in press)) used (Mazaud *et al.*, 1991; McElhinny and Senanayake, 1982; Meynadier *et al.*, 1992), their calculations still illustrate the substantial effects field strength changes will have on *in situ* cosmogenic ages. Clapp and Bierman (1995) have recently converted the effects of geomagnetic field-strength fluctuations on specific production rate-calibrations into a temporally continuous production rate curve for the past 140 ky. The model, reproduced as a computer program, calculates exposure-ages as a function of the measured isotope abundance, latitude, and altitude of a sample site, along the lines of the radiocarbon calibration program CALIB (Stuiver and Reimer, 1993).

IMPLICATIONS

Interpreting measured isotope abundances as time-averaged production rates and applying these production rates to sites at different altitudes and latitudes requires assuming or estimating a variety of parameters including: the exposure age of the calibration surface, the time-averaged geomagnetic latitude of the sample sites, corrections for exposure geometry, scaling factors for elevation and latitude, and the efficiency of various isotope production pathways. Varying these parameters can generate significant differences in production rates (Table 4.2). For example: 1) Nishiizumi *et al.* (1989) assume that the current geomagnetic latitude of their calibration samples (44°) is representative of the average geomagnetic latitude since Tioga deglaciation. However, if the average geomagnetic latitude since exposure is taken as the current geographic latitude (38°), as was done for Meteor Crater data (Nishiizumi *et al.*,

Table 4.2: ^{10}Be Production Rates (atom g^{-1} yr^{-1}) at Sea Level, High Latitude

Exposure age (cal kyr)	Location	*Geomagnetic latitude during exposure*					Sample elevation (km)
		****44°***	****38°***	***†44°***	***†38°***	***†41°***	
11	Tioga, Sierra Nevada	6.03	6.85	5.36	6.09		2.1-3.5
13	Tioga, Sierra Nevada	5.10	5.80	4.53	5.16		2.1-3.5
14	Tioga, Sierra Nevada	4.74	5.39	4.21	4.79		2.1-3.5
21.5	Laurentide, NJ					4.76	0.2-0.3

* production rate scaled by Nishiizumi *et al.* (1989) assuming muon production as calculated by Lal and Arnold (1985)

† production rate scaled assuming all production by spallation

1991) and as is indicated by measurements of late-Holocene shifts in the geomagnetic dipole (e.g., Merrill and McElhinny, 1983, p. 100), calculated high latitude/sea level production rates would be 15% higher than currently reported. In effect, such a difference in integrated paleomagnetic latitude for the Sierran calibration sites would have the same result as a 15% overall decrease in integrated paleomagnetic field strength. 2) Using 14,000 cal yr B.P. rather than 11,000 cal yr B.P. for the exposure age of the Sierran calibration samples, as seems reasonable from my study, lowers all production rates by more than 20%.

The ^{10}Be production rate Larsen and Bierman estimate from the Laurentide data (4.76 atom g^{-1} yr^{-1}, assuming a 21,500 yr exposure age) is most consistent with Nishiizumi *et al.*'s (1989) data if Sierran deglaciation occurred between 13,000 and 14,000 cal yr B.P. regardless of which geomagnetic latitude is used (Clark, 1995 (in press)). Paleointensity-based calculations can also be used to support the conclusion that the calibration sites of Nishiizumi *et al.* (1989) were deglaciated before 11,000 cal yr B.P. Data from Meteor Crater (Nishiizumi *et al.*, 1991) are consistent with longer-term average sea-level, high-latitude ^{10}Be and ^{26}Al production rates of about 6 and 36 atoms g^{-1} (Nishiizumi *et al.*, 1989). Using the paleomagnetic intensity record of Clark *et al.* (1995 (in press)), it is difficult to reconcile the 11,000 cal yr B.P. Sierran calibration with the Meteor Crater data. Assuming an average geomagnetic latitude of 38° for the Sierran calibration samples and assuming that the 11,000 cal yr B.P. calibration and the paleointensity record were both correct, Meteor Crater should appear to be about 58,000 yr B.P. However, Meteor Crater has been independently dated by thermoluminescence at 49,000 ± 3000 yr B.P. (Sutton, 1985); it is more likely, therefore, that the 11,000 cal yr B.P. calibration age is in error. Although perhaps not obtained from the most reliable dating method, the thermoluminescent age is the only independent measure of the age of the impact. If one accepts the luminescence age and applies a production rate calculated from the Meteor Crater data, and uses the paleointensity-based forcing proposed by Clark *et al.* (1995 (in press)), the exposure age of the Tioga calibration surfaces would be about 12,900 cal yr B.P., independently supporting the new radiocarbon data presented above.

CONCLUSIONS TO CHAPTER 4

The new radiocarbon ages for Sierran deglaciation, along with the Laurentide ^{10}Be data and the calculations of the effects of paleomagnetic field fluctuations, and other significant geologic uncertainties (Bierman and Gillespie, 1991; Hallet and Putkonen, 1994) show that age estimates based on the abundance of *in situ* produced cosmogenic isotopes are not yet certain enough to warrant their use in correlating brief (500 -1000 yr) geomorphic or climatic events, such as the late-Pleistocene Younger Dryas climatic reversal or Heinrich events.

The revised chronology of deglaciation in the Sierra Nevada emphasizes the need for accurate, independent age control at all calibration sites. Exposure ages accurate to within several percent are required to define accurately short-lived events or to correlate reliably cosmogenic ages with other well-dated time series. To achieve this level of both accuracy and precision, several other parameters will also need to be better understood, including: the effective long-term geomagnetic latitude of a sample site, altitude and latitude scaling corrections, and the relative contribution of muons to ^{10}Be and ^{26}Al production. In addition, instantaneous production-rate corrections, such as demonstrated in Clark *et al.* (1995 (in press)), will need to be refined to take into account the changing intensity of the geomagnetic field.

Cosmogenic exposure ages will be more accurate and useful if independent production-rate calibrations are performed at a variety of sites spanning a range of altitudes, latitudes, and durations. Calculations show that errors in relative and numerical ages, related to the varying intensity of the geomagnetic field, will be minimized if calibrations are made and samples are collected at high latitudes and low altitudes. Errors related to applying locally derived production rates to distant sampling sites will decrease if the altitude and latitude of sample and calibration sites are similar. Finally, errors related to geomagnetic forcing will be reduced if samples and calibration sites are exposed for similar durations.

SUMMARY

To summarize, the major findings of my doctoral research include:

1. Abundant glacial deposits in the northern Sierra Nevada indicate that several large, previously undocumented ice fields and ice caps covered significant portions of the crest during the late-Wisconsin Tioga stade. The largest ice fields occupied the headwaters of the Mokelumne River, the Truckee/North Fork American rivers, North and South Yuba rivers, and the North Fork Feather River. In addition, there is strong evidence for numerous valley glaciers that occupied north- and northeast-facing cirques and ridges between and adjacent to the larger ice fields. ELA estimates based on highest occurrences of lateral moraines from the ice fields conform with Tioga ELAs estimated in the southern half of the Sierra Nevada (Burbank, 1991; Gillespie, 1991), and indicate that the northward latitudinal snowline gradient decreased by about 2.5 m/km, while the cross-range gradient climbed by about 5.8 m/km to the east. These results mimic modern temperature/precipitation gradients in the Sierra Nevada, suggesting that the late Wisconsin glacial climate resulted from an intensification of the modern regional climate pattern. Although locally there are distinct pre-late Wisconsin moraines in the northern Sierra Nevada, they are either not as ubiquitous or not as well preserved in the southern end of the range.

2. Post-Tioga cirque deposits in the southern Sierra Nevada represent two separate advances: the Recess Peak, and the Matthes. Exposures of ice in the voluminous Matthes moraines and in related valley-floor rock glaciers indicate that they have large ice cores that are covered by only a thin (1-10 m) veneer of rockfall derived from the cirque headwall. Despite a number of recent studies that argue vehemently that all rock glaciers are periglacial in origin, my data strongly support a glacial origin for valley-floor rock glaciers in the Sierra Nevada, placing them in a continuum encompassing clean

glaciers, ice-cored moraines, and rock glaciers. Valley-wall rock glaciers, which typically originate from avalanche chutes and scree slopes and flow only short distances towards the valley floor, may more closely resemble the concept of "periglacial" rock glaciers, although the likely primary mode of ice accumulation (via snow-rock avalanches) is also important in feeding true cirque glaciers.

3. The mantle of debris on valley-floor rock glaciers acts as an effective insulator that inhibits ablation. As a result, such debris-covered glaciers become larger and reach to substantially lower altitudes than do adjacent bare-ice glaciers with equivalent accumulation areas. Consequently, the accumulation area ratios (AARs) and toe-to-headwall altitude ratios (THARs) of such glaciers, commonly used to estimate ELAs for extinct glaciers, are substantially different (0.05-0.2 and 0.6-0.8, respectively) than for clean glaciers (~0.65 and 0.4-0.5, respectively). Debris-covered glaciers were also common in the southern Sierra Nevada during the Recess Peak advance. Reconstructed ELAs for these glaciers using an AAR of 0.15 are up to several hundred meters higher than they are if the effect of the debris cover is not accommodated. With debris-cover taken into account, Recess Peak ELAs along the Sierran crest form a southward-rising gradient that is equivalent to those of the Tioga, Matthes, and modern glaciers, and which represents a snowline lowering of about 20% of full glacial conditions. Although less common than for Recess Peak or Matthes advances, debris -covered glaciers occupied at least several drainages east of the Sierran crest during the Tioga stade. ELA studies that do not distinguish former debris-covered glaciers in regions where they occur will introduce substantial error into their reconstructions.

4. Age of the Recess Peak advance has been controversial; since first recognized, it has generally been assigned to the mid- to late-Neoglacial period, roughly 2-3000 yr B.P., based primarily on relative weathering parameters and lichenometry. Previous work indicates that the maximum Matthes advance postdates ~650 cal yr. B.P. Basal

radiocarbon dates from cores of four high-altitude lakes dammed behind Recess Peak moraines in three locations show that the advance predates the Neoglacial period, and dates from the southernmost coring site in Bishop Creek demonstrate that the Recess Peak advance was over before ~11,200 ± 70 ^{14}C yr B.P. (~13,100 cal yr B.P.). The latter minimum-limiting date indicates that the Recess Peak advance predates the North Atlantic Younger Dryas cooling event; absence of any moraines between those of the Recess Peak and Matthes advances demonstrates that there were no glacier advances larger than the Matthes during the ~12,000 yrs that separate the two advances, including any during the Younger Dryas and early Neoglacial periods.

5. Prior to this study, it was widely assumed by glacial geologists that the Sierra Nevada was deglaciated about 10,000 yrs ago. The new minimum dates for Recess Peak deglaciation, and basal dates from a number of other post-Tioga lakes and meadows, indicate that deglaciation in the Sierra Nevada began by 18,000-19,000 cal yr B.P., and was completed before 13,000 cal yr B.P. The latter date is particularly important for anyone interested in surface exposure dates using in-situ produced cosmogenic ^{10}Be and ^{26}Al, because the late-Pleistocene production rates for these isotopes were estimated from glacially polished Sierran bedrock samples that were presumed to be deglaciated 11,000 cal yr B.P. The samples used for calibration all occur downstream of Recess Peak deposits in those drainages; therefore, the sites had to have been exposed before 13,000 cal yr B.P. The 2000+ yr discrepancy between the revised and original calibrations introduces a systematic error of at least 19% for late-glacial samples dated using the current production rates for these isotopes. The earlier age of deglaciation of the range is supported by independent production-rate measurements on rocks from late-Wisconsin Laurentide moraines in New Jersey, and by calculations of the effects of geomagnetic field fluctuations on the long-term production rates of these isotopes. Together, these findings indicate the need for caution when interpreting surface exposure dates. The

analytic errors (~5%), which typically have been the only ones reported for such dates, are clearly insufficient when the geologic uncertainties inherent to the original calibrations are considered.

In addition to these new findings on the glacial record of the Sierra Nevada, many important problems remain. For example, the nature, spatial patter, and timing of Tioga deglaciation remains unclear. Included in this problem is the need for a maximum limiting age for the Recess Peak advance: was the Recess Peak event merely a last-gasp readvance of the Tioga glaciers? Or was it a wholly separate event that followed complete deglaciation of the range? Following the same line of thought, a full characterization of the timing and extent of glacierization in the Sierra Nevada during the Holocene (i.e., Neoglaciation, Little Ice Age) remains unrealized. Further investigation into the internal structure, genetic processes, and kinematics of Sierran rock glaciers is needed. And finally, detailed studies of glacial deposits in the northern end of the range are needed, especially to identify the extent and character of pre-Tioga deposits. Because of the abundance of vegetation and different rock types in this part of the range, there may be opportunities to date moraines directly,. e.g., via radiocarbon or weathering rinds, which do not exist further to the south. In a more regional view, my Recess Peak results beg the question whether a correlative event affected other mountains in western North America. There is support for a Younger Dryas advance in the Canadian Rockies, and for both late-glacial and early Holocene (?) events on Mt. Rainier (J. Heine, personal communication, 1995). How are these events expressed in the mountains of the southern Cascades, Rocky Mountains, and Great Basin? For that matter, how do they relate to the glacial record in the North Pacific, Hawaii, and Asia, because they are directly "upwind" of western North America. I hope to address these and related questions in my future work.

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APPENDIX A: COMMENT ON THE NATURE OF ROCK GLACIERS IN THE SIERRA NEVADA

The following text is excerpted from a comment by M. Jakob (1994) concerning my work, and that of my co-authors, on rock glaciers in the Sierra Nevada (Clark *et al.*, 1994a). It, and our response (Clark *et al.*, 1994b), present succinctly the two primary, and contradictory, views on rock glacier genesis, and I think it helps reemphasize some points on the origin and nature of these landforms. It also highlights the real need for clarification of terminology in this field. Many of Jakob's criticisms of our findings stem from his making a hard distinction between what is "glacial" and what is "periglacial," when much of the distinction is artificial, and a product of lexicon. If process is the primary consideration, there is much overlap between these two groups.

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LETTER TO THE EDITOR

Comments on "Debris-Covered Glaciers in the Sierra Nevada, California, and Their Implications for Snowline Reconstruction," by D.H. Clark, M.M. Clark and A.R. Gillespie

In Quaternary Research, vol. 41, No. 2, Clark et al. (1994) discuss rock glaciers and their implications for snowline reconstruction in the Sierra Nevada, California. In the first part of their paper the authors claim that the rock glaciers studied are actually debris covered glaciers of late Holocene age. In the second part of the paper they recalculate equilibrium line altitudes (LIAs) according to their findings. In this letter I challenge both their interpretation of rock glaciers as debris covered glaciers and accordingly their conclusion regarding snowline reconstruction. The evidence presented strongly suggests that the rock glaciers in question are indeed of periglacial origin, and indicate the existence of at least discontinuous alpine permafrost in their area.

Clark et al. (1994:139-153) oppose the widely held view that rock glaciers are a morphological expression of mountain permafrost (e.g. Barsch 1987, 1988, 1992, Cui and Zhu Cheng 1988, Giardino and Vitek 1988, Gorbunov 1983, Haeberli 1985, Haeberli and Schmid 1988, Haeberli et al. 1988, Haeberli et al. 1992, Keller and Gubler 1993, Vonder Mühll and Haeberli 1990, Vonder Mühll and Schmid 1993) and quote several authors who have "... demonstrated that some rock glaciers are cored by glacier ice mantled with a continous and relatively thin debris cover" (page 139). None of these authors has presented enough evidence to unambigously show the existence of a continuous ice core of glacial origin.The periglacial model of rock glacier origin is based on a large number of ground temperature measurements, electrical resistivity, radio-echo and seismic refraction soundings as well as core drilling through an active rock glacier in the Swiss Alps (Haeberli et al. 1988). These measurements were not only carried out in the Alps but also in the Argentine Andes (Ahumada 1992, Happoldt and Schrott, 1992, Schrott 1991), the Himalayas (Jakob 1992, Barsch and Jakob 1993, Fujii and Higuchi 1976) and the mountains of mainland Norway and Svalbard (Sollid and Sørbel 1992). These observations affirm that the elevation of many rock glacier termini coincide closely with the lower limits of discontinous alpine permafrost in the representative areas.

The absence of contemporary permafrost features, as noted by the authors, cannot be used as an argument for the non-existence of permafrost. Morphologic, textural and hydrologic conditions in steep alpine terrain rarely allow the development of morphological expressions of permafrost other than rock glaciers themselves. Even if some of these forms are found they do not necessarily indicate permafrost conditions. For example stone stripes or small stone circles can develop from diurnal freeze-thaw cycles (e.g. Williams and Smith 1989). Since Clark et al. question rock glaciers as indicators of mountain permafrost, ground temperature measurements over a whole year would be helpful in clarifying this question.

On page 141 Clark et al. note that glacier ice on Southfork Pass Rock glacier continues beneath the debris cover and does not end there. They present a photograph of a meltwater pond in the upper reaches of the rock glacier as proof (page 143, fig. 3). Without further evidence the claimed continuity of glacial ice must be placed in doubt. The meltwater depression is called a "thermokarst pond" by the authors. By definition, thermokarst requires the presence of degrading permafrost (Harris et al. 1988, Williams and Smith 1989). However, the pond is surrounded by overhanging ice. The ice is identified as glacial in origin from the presence of silt layers, which are separated by zones of clear ice thought to represent annual layering. Horizontal layering of ice-rich silt and sand is a common phenomenon in permafrost environments (e.g. Mackay and Dallimore 1992 fig. 4, Haeberli et al 1988 fig. 5), and therefore cannot used alone to determine the origin of an ice body. Ice fabrics, ice bubble patterns and structural details could be examined to help understand the origin of the ice. In addition, water quality (i.e. ionic concentrations and oxygen isotopes) of the rock glacier ice could be compared with that of nearby glacier ice. Even if the observed ice is glacial in origin, this does not allow the conclusion that the rock glaciers are indeed ice cored throughout their entire length, since dead glacier ice can be incorporated into rock glaciers (Haeberli 1989).

Several of the authors' own observations point to permafrost conditions and a periglacial origin for the rock glaciers in the study area. Clark et al. list extensive ablation drift upslope of the rock glacier termini, residual longitudinal and transverse ridges in the drift, thermokarst features, and a concave deflated character of the debis mass as evidence for remnant debris covered glaciers (p. 147). Debris covered glaciers, however, typically do not display longitudinal or transverse ridges but lose their flow morphology as they downwaste. This, and the other criteria, point toward a typical rock glacier.

On page 140, the authors differentiate between lobate and togue shaped rock glaciers - a widely used morphological distinction. They write: "... (in contrast to valley wall [*lobate*] rock glaciers which may be periglacial in origin)." The morphological expression does not necessarily imply different internal structure, and the presence or absence of permafrost. Clark et al. indicate a periglacial origin by mentioning internal shearing along ice lenses within the rock debris and the existence of local ice within the creeping talus rock mass [*lobate rock glaciers*]. The authors state on page 144 that "...valley wall rock glaciers reach altitudes more than 200 m below the lowest contemporary debris-covered glaciers...", contradicting their conclusion that permafrost is absent from the area (page 143). The conclusion seems to be that if those lobate rock glaciers are still active, permafrost does exist even below the lower limit of the tongue shaped rock glaciers. If so, this would suggest a lower limit of discontinous alpine permafrost at around 3200 m in their study area.

A one meter pit was excavated by the authors into the steep face of the rock glacier terminus. Saturated silty matrix was encountered at a temperature of "...only slightly above 0 dgrees C" (page 141). This observation, which was carried out in late summer, probably indicated the existence of a one-meter thick active layer with permafrost below. In late summer and early fall, positive temperatures can prevail to a depth of several meters, below which permafrost is encountered as demonstated by Haeberli et al (1988) on the rock glacier Murtèl I in the Swiss Alps.

On page 143 the authors estimate rock glacier creep (flow) rates at 10-20 cm per year based on photogrammetric measurements of the movements of an indiviudual large boulder. Linear extrapolation of these rates using the length of the Southfork Pass Rock Glacier of 1600 m would yield ages of 8000 - 16,000 years. These ages are an indication that the rock glacier commenced its movement at the end of the last glaciation. Clark et al., however, assume that the age of the rock glacier coincides with the end of the late-Holocene Recess Peak advance, without explaining the discrepancy between their assumption and the extrapolated age.

On page 144-146, Clark et al. describe the insulating properties of debris covers on glaciers, quoting the Balch ventilation, and concur that the temperature within the debris cover is close to freezing even in summer. These low temperatures constitute another strong argument for the occurrence of permafrost in the region.

The second part of the paper is based on the assumption that the rock glaciers are remnants of debris covered glaciers of Recess Peak age, contradicting the view of Burbank (1991) and Yount et al. (1982). Without presenting absolute dates or more conclusive evidence, this assumption and further implications regarding snowline elevations, cannot be supported.

The periglacial model of rock glacier formation has been challenged and critizised many times in the past. However, the evidence presented fails to prove a glacial origin for the rock glaciers, and in some cases even points towards a periglacial origin. Permafrost describes the thermal condition of the ground which is crucial for the long-term survival of underground ice. It cannot be assessed without the appropriate measuring techniques. The following additional observations would greatly clarify the origin of their rock glaciers:

- year-round ground temperature measurements at different altitudes and different aspects to investigate possible permafrost conditions

- absolute ages of the rock glaciers and (other) Recess Peak deposits

- geophysical measurements on nearby glacier ice and rock glaciers to differentiate between different ice characteristics and to detect an active layer if present

- ice cystallography of several samples taken at various locations on, and at various depths within the rock glacier to differentiate between glacial and non-glacial ice.

In summary, the evidence presented in the paper is insufficient to reject the periglacial origin of the tongue-shaped rock glaciers, and also is insufficient to demonstrate a continuous ice core within the rock glaciers. Contrary to the authors views, it seems likely that discontinuous permafrost, with a lower limit between 3200 and 3400 m, is common in their area. Finally, if the rock glaciers developed during and after Pleistocene deglaciation, as indicated by Burbank (1991) and Yount et al. (1982), and not during the Recess Peak Neoglacial advance, as tentatively confirmed by linear extrapolation of present creep rates, the recomputation of equilibrium line altitudes and their paleo-climatic implications must be questioned.

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Curriculum [redacted]

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RESEARCH INTERESTS:

Paleoclimatic significance of alpine and lacustrine deposits; geomorphic evolution of mountain regions; tectonic development of western North America. and central Asia.

EDUCATION:

Ph.D., 1995, Geology, University of Washington
Dissertation: *"Extent, timing, and climatic significance of latest Pleistocene and Holocene glaciation in the Sierra Nevada, California"*

M.S., 1986, Geology, Stanford University
Thesis: *"Tectonic evolution of the Red Hills-southwestern Kern Mountains area, east-central Nevada"*

B.S., (with Distinction), 1983, Geology, Stanford University

ACADEMIC EXPERIENCE:

1989 - 1995 Doctoral Candidate/Lab Instructor/Research Assistant (Dept. Geol. Sciences, Univ. of Washington): developed and taught introductory geology labs and field trips; mapped and analyzed climatic significance of geomorphic surfaces in western U.S. and central Asia.

1984, 1985, 1990 Field Mapping Instructor (Stanford U., Univ. Wash.): helped develop, teach, and evaluate advanced field mapping and sampling methods for upper-level geology majors.

1983 - 1985 Graduate Student/Lab Instructor/Research Assistant (Dept. of Geology, Stanford University): developed and taught introductory geology labs and field trips; mapped and analyzed supracrustal structures in the eastern Basin and Range.

PROFESSIONAL EXPERIENCE:

1996 Research Associate (Geological Remote Sensing Laboratory, Univ. of Washington): Investigate climatic and geomorphic significance of alpine glacial and lacustrine deposits in western U.S. and central Asia, including extensive use of cosmogenic nuclides for exposure ages and landform evolution.

1986 - 1989 Geologist (Earth Sciences Associates, Palo Alto, California): mapped and analyzed surficial deposits and crustal structures for seismic and land-failure hazards.

GRANTS, SCHOLARSHIPS, AND AWARDS:

D.O.E. Global Change Fellowship (University of Washington, 1991-95)
Harry Wheeler Scholarship (Univ. of Washington, 1993)
George E. Goodespeed Fellowship (Univ. of Washington, 1992)
White Mountain Research Station Research Grant (Univ. of Washington, 1992)
Richard E. Fuller Scholarship (Univ. of Washington, 1991)
GSA Research Award (Univ. of Washington, 1990, 1991, 1992)
Outstanding Mention, 1992
Sigma Xi Research Award (Univ. of Washington, 1990,1992)
Departmental Research Award (U. of Washington, 1990, 1991, 1992, 1993)
Fink Graduate Award (Stanford University, 1983-1985)
Athletic Scholarship, Varsity Soccer (Stanford University, 1979-1983)
Co-captain, MVP, 1982

OTHER EXPERIENCE:

Conceived and convened conferences on glacial history of the Sierra Nevada (9/95, White Mountain Research Station, Bishop) and an AGU Chapman Conference on the climatic significance of rock glaciers (scheduled for 8/96); Lab-setup, sampling, and analyses of lacustrine and surface deposits (^{14}C, cosmogenic nuclides), and bedrock structures (cosmogenic nuclides, $^{40}Ar/^{39}Ar$, conodont-color paleothermometry, fission-track dating); Trench and borehole logging.

PROFESSIONAL AFFILIATIONS:

American Quaternary Association
Geologic Society of America
American Geophysical Union

PUBLICATIONS:

Papers

Clark, D. H., and Gillespie, A. R., 1995 (in press), Timing and significance of late-glacial and Holocene glaciation in the Sierra Nevada, California. *Quaternary International.*

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