Rangewide glaciation in the Sierra Nevada, California

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ABSTRACT

The 600-km-long Sierra Nevada underwent extensive Pleistocene glaciation except for its southernmost 100 km. Presently, ~1700 small glaciers and ice masses near the crest of the range occur above 3250 m in elevation; these covered an area of ~50 km² in 1972. Fourteen of the largest glaciers decreased by about one half in area during the period from 1900 to 2004.

Rock glaciers, generally glacial ice covered by 1–10 m of rockfall debris, occur in about the same span of the range as ice and permanent snowfields. They are, on average, lower by 200–300 m, apparently because of the insulating layer of rocky rubble that protects their internal ice from the sun’s heat and from wind.

The principal Pleistocene glacial stages are the Sherwin (ca. 820 ka), Tahoe (130–170 ka), and Recess Peak (13 ka). Some 7040 glacial lakes, produced primarily by quarrying from bedrock, were mostly exposed after recession of the Tioga glacial stage. The lakes largely mark the area of primary snow accumulation. Below the lower limit of the lakes, ice flowed downward into river-cut canyons, forming major trunk glaciers within the zone of ablation.

The range is in general a westward-titled block upfaulted on its east side. Therefore, the main late Pleistocene trunk glaciers (Tahoe/Tioga) west of the crest extend 25–60 km, whereas those east of the crest extend only 5–20 km. Because of higher precipitation northward, glacial features such as the toes of existing glaciers and rock glaciers, as well as the late season present-day snowline, all decrease in elevation northward. Likewise, the elevation of the lower limit of glacial lakes, an indication of the zone of snow accumulation during the late Pleistocene, decreases about the same degree. This similarity suggests that the overall climate patterns of the late Pleistocene, though cooler, were similar to those of today. The east slope glaciers show a similar northward depression, but they are ~500–1000 m higher.

The upper part of the glacial system was erosive over a broad highland area as the evenly distributed ice in the accumulation zone moved to lower elevation. The abundant lake basins record this erosive action. The lower part of the glacier system was largely confined to major preexisting river canyons in which melting dominated. The average of rangewide estimates of the equilibrium line altitude (ELA)—the boundary between the upper snow and ice accumulation zone and the lower ablation zone—of many late Pleistocene glaciers parallels, and is only 200–300 m above, the altitude of the lower limit of the lakes. Hence, the lake zone provides a means of estimating the ELA.

INTRODUCTION

The north-northwest–trending Sierra Nevada of eastern California is 600 km long (430 km north-south) and has a crest elevation from 2000 to 4400 m. Presently, more than 1000 small glaciers and permanent ice masses occur near the crest of the range. They represent a tiny part of the area covered by ice during the Pleistocene glacial stages, when 490 km of the crest and upper region of the range were glaciated. Glaciation strongly modified the topography of the higher terrain, producing sharp peaks, steep canyons, moraines, and more than 7000 lakes (Figs. 1 and 2).

We compiled quantitative data on the more than 1000 present-day glaciers and ice masses, as well as on rock glaciers, to compare with present-day snowpack history and rangewide temperature and precipitation trends in order to judge conditions of climate change. These data were then compared with a new compilation of the extent of the ice over the entire range during the late Pleistocene. Some of these glaciers spread out on the east piedmont slope of the range, but the main trunk glaciers on the west side descended only part way down previous river canyons, which they modified into U-shaped canyons, such as Yosemite Valley.

Many detailed studies have examined the Pleistocene glacial history of selected regions, particularly in the better-exposed east side of the range. However, in this work, we attempt an integrated view of the late Pleistocene glaciation of the entire range. The limit of such glaciers on both sides of the range has been compiled, providing data to calculate the equilibrium line altitudes (ELA) for all the major glaciers. These ELAs descend north at ~3 m per kilometer of latitude and are systematically higher on the east side of the range. In this regard, they mimic the present-day May snowline and also the lower limit of the zone of the 7040 glacial lakes, which formed in the late Pleistocene glacial accumulation zone. This correspondence provides insight into the climate during Pleistocene glaciation as well as an independent method of estimating the ELA.

DATA SETS

Elevations were derived from the U.S. Geological Survey (USGS) 10 m National Elevation Data set (NED). Location (centroid) and size of glacial lakes, permanent ice, and areas covered by timber are from 1:24,000 scale Digital Line Graphic (DLG) analysis of USGS topographic maps. A 1991 field survey of ice, based in part on 300 aerial photographs taken by Austin Post in 1972, provided supplementary data on present-day glaciers and permanent snowfields (Raub et al., 2006).

Rangewide temperature and precipitation data were acquired online from the Prism Climate Group (2004; Daly et al., 1994). Seasonal snow cover on the range is available from the National Aeronautics and Space Administration (NASA) satellite with MODIS (MODerate-resolution Imaging Spectroradiometer) (Hall et al., 2002; Dozier et al., 2008).

Three regional studies show the extent of late Pleistocene glaciation: in Yosemite National Park (Matthes, 1930; Alpha et al., 1987), in the drainage of the San Joaquin River (Matthes, 1960), and in Sequoia and Kings Canyon National Parks (Moore and Mack, 2008). Only a few studies provide detailed information on...
Figure 1. Map of eastern California showing topography of the Sierra Nevada with glacial lakes shown by blue dots. The lakes were exposed and formed at the retreat of the final Tioga glacial stage. Note that the elevation of the range and the width, elevation, and density of the lake zone decrease toward the north.
The limit of glaciation in the west slope canyons north of Yosemite, because morainal deposits have largely been removed or are poorly preserved due to river erosion. We used topographic evidence to estimate the extent of many of the western trunk glaciers.

Glacial moraines and rock glaciers have been mapped during geologic studies that cover a large part of the range. Most of these maps are included in the series of 15 min USGS geologic quadrangle maps compiled and duplicated online in the Sierra Nevada Batholith Mosaic Project (Caudill, 2005). Moraine information is also found on the half-degree quadrangle maps of the USGS geologic folios (1:250,000) published around the beginning of the twentieth century, and on the more modern 1° × 2° maps of the USGS geologic folios (1:125,000) cited to E. Blackwelder, and the south half (dated 1937) is credited to F.E. Matthes. This map, and later small-scale maps showing the extent of glaciers over the entire range, indicates that the range crest was largely covered by a continuous ice cap (Wahrhaftig and Birman, 1966). Actually, considerable areas near the crest, particularly ridges and peaks, were ice free, as shown in the more detailed maps of Yosemite National Park (Alpha et al., 1987), and Sequoia and Kings Canyon National Parks (Moore and Mack, 2008).

Blackwelder (1931) divided Pleistocene glaciation into four stages: McGee, Sherwin, Tahoe, and Tioga, from oldest to youngest. Various modifications and additions to this scheme have been made since, but these glacial stages still form the basis for modern studies.

The first radiometric constraints on the age of glacial deposits became available with K-Ar dating of volcanic rocks associated with moraines (Dalrymple, 1964). More recently, the technique of surface exposure dating, which measures the accumulation of cosmogenic nuclides, principally $^{36}$Cl (Phillips et al., 2009) and $^{10}$Be (e.g., Rood et al., 2011), has become a valuable tool for dating morainal material.

**FEATURES OF THE RANGE**

**Topography**

The 600-km-long Sierra Nevada trends 30° west of north and varies in width from 80 to 110 km (Fig. 1). Overall, the range is a coherent physiographic block tilted west, and it is delimited on much of its east slope by normal fault-line scarps, which commonly define a steep eastern escarpment, contrasting with a much gentler western flank. From its southern end at the Garlock fault, the mountain crest rises northward for 145 km to its highest point at Mount Whitney (4418 m) at 36.58°N latitude (Fig. 2).

Continuing north from Mount Whitney, the crest maintains an elevation of greater than 3200 m for 140 km to the Mammoth region near the head of the San Joaquin River. A decrease in general elevation occurs in this region, but the >3200 m crestal region continues north for an additional 107 km to 38.75°N latitude near the head of the Stanislaus and Mokelumne Rivers.
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From there, the crest decreases for 150 km to ~2200 m at its north limit, south of the northern forks of the Feather River at 39.75°N latitude.

North-trending normal faults along the east side of the range mark the western limit of the Basin and Range Province, a region characterized by crustal extension. The first major fault block east of the southern Sierra Nevada is the White-Inyo Mountain Range, which—though nearly as high—is only slightly glaciated because it lies in the rain shadow of the Sierra Nevada. East of the central Sierra Nevada, there are the slightly glaciated Sweetwater Mountains. East of Lake Tahoe, in the northern Sierra Nevada, there is the Carson Range, which is likewise only slightly glaciated. These three eastern outliers of the range are excluded in this study.

Major rivers have cut giant canyons on the west side of the range. These canyons and their tributaries have produced large moderate-elevation reentrants in the west slope (Fig. 1). The upper parts of these river canyons were occupied by trunk glaciers that carved giant U-shaped canyons.

Climate

Rain and snowfall increase markedly in the Sierra Nevada from south to north. Annual precipitation at the range crest is highly variable from year to year, but the 30 yr average is 800 mm in the south, and this increases systematically to 1300 mm in the north. Likewise, the 30 yr average precipitation at 2000 m elevation on the west slope increases about the same amount from south to north (Prism Climate Group, 2004; Fig. 3). This increase is apparently caused by the overall northward concentration of storm tracks commonly diverted from the south by the prevalence of elevated barometric pressure zones. The orographic effect of the high range causes heavier precipitation and lower snowline on the west slope compared to the steep east slope (Fig. 4).

The 30 yr average (1971–2000) of daily maximum and minimum temperatures at the range crest are warmest at the south end, at 15 °C and 2 °C, respectively (Prism Climate Group, 2004; Fig. 3). Proceeding north, they are both coolest in the middle high part of the range at 5° and ~5 °C and then increase to 12 °C and 1 °C, respectively, in the north. In contrast, the 30 yr average maximum and minimum temperatures at 2000 m elevation decrease systematically from 17 °C and 5 °C, respectively, in the south to 14 °C and 1 °C, respectively, in the north. Present-day ice occurs along the range crest, where the 30 yr average minimum temperature at the range crest is ~3 °C to ~5 °C (Fig. 3).

Snow covers much of the range in the winter and spring, but the snowpack melts back in the summer and fall. NASA’s satellite with MODIS provides useful images of the extent of the snowpack (Fig. 5), which begins growing in the winter and starts shrinking in February-March to finally disappear in August-September. The extent of the snowpack varies considerably from year to year, such as 2006. The average elevation of the snowline on the west slope of the range (Fig. 4A) is somewhat irregular up through February because of the accumulation and melting of snow dumped by individual localized storms. However, through April and May, the altitude of the snowline systematically rises from 1500 to 2600 m at 38°N latitude as temperatures increase and the snow melts.

In contrast, the east side snowline (Fig. 4B) is systematically higher by roughly 500–1000 m compared to that of the west. It follows a similar pattern of rising through April and May, but it moves from 2400 to 3200 m at 38°N latitude. The base of the snowpack on both west and east sides tilts down to the north at midseason at ~1.3–2.0 m/km of latitude (Fig. 4), primarily because of higher precipitation northward.

During average years, snowpack width at 38°N lat is ~52 km (Fig. 6). An example of the growth and shrinkage of the snowpack is shown by the position of its lower limit (the snowline) during an average year, such as 2006. The average elevation of the snowline on the west slope of the range (Fig. 4A) is somewhat irregular up through February because of the accumulation and melting of snow dumped by individual localized storms. However, through April and May, the altitude of the snowline systematically rises from 1500 to 2600 m at 38°N latitude as temperatures increase and the snow melts.

Figure 3. Climate data for the Sierra Nevada for 1971–2000, showing latitude range of present glaciers: (A) temperature (°C) and (B) precipitation (mm) (from Prism Climate Group, 2004).
The late season snowpack has evened out the variations due to individual storms of the early season. Aside from the overall precipitation, the snowpack is controlled largely by the conditions of meltback. The decrease of the snowpack elevation northward in the range, its higher elevation on the east side, and its overall shape in the spring and early summer all mimic the area covered by glacial lakes. This likeness indicates that the late Pleistocene climate patterns, though colder, were similar to those of today and that the present-day snowpack can serve as a template for the much larger and longer-lived snowpacks that accumulated during the glacial stages.

**Vegetation**

Much of the range is heavily forested, and only in the highest 200 km (between 36.5°N and 38.2°N lat) is the crest nearly devoid of trees (Fig. 2). Timberline decreases systematically from ~3700 m in the Mount Whitney region northward to 3500 m at 38.2°N latitude, where timber extends to the range crest. North of this, the rest of the range is totally forested (Fig. 2). The present-day giant sequoia groves grow on the west slope from 1500 to 2200 m in elevation, 700–1500 m below timberline (Fig. 7). In general, forests are thicker on the west slope. The sequoias extend above the lower extent of late Pleistocene glaciers, but between and below current extensively glaciated regions.

**PRESENT GLACIATION**

**Present-Day Ice**

U.S. Geological Survey topographic maps made from aerial photos taken in 1976–1978 show 1784 discrete areas of ice that include glaciers and permanent snowfields that are identified by blue topographic contours. In this data set, the discrimination of glaciers from snowfields and late season snow can reflect the judgment of individual topographers and may not be uniform from area to area. A separate inventory derived from 300 oblique aerial photographs taken in August 1972, during a particularly dry year, identified 497 glaciers and 788 ice patches (total 1285), which were tabulated and plotted on USGS 15 min topographic maps (Raub et al., 2006). These two data sets show a similar, though slightly different, distribution of ice.

The ice masses occur near the crest in the highest part of the range over a distance of 280 km from 36.3°N lat to 38.6°N latitude (Fig. 8). The ice occurs in that part of the range where the crestal 30 yr average minimum daily temperature is −3 °C to −5 °C (Fig. 3). The mapped ice

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**Figure 4.** Elevation of snowline on the (A) west and (B) east sides of snowpack from February to June in 2006, an average snow year. From the National Aeronautics and Space Administration (NASA) satellite with MODIS (MODerate-resolution Imaging Spectroradiometer). The snowline remains higher through the year on the east slope primarily due to the orographic effect on storm clouds from the Pacific Ocean.

**Figure 5.** Extent of snowpack during 2006 (an average snow year) for the beginning of March, April, May, and June. Blue margins indicate areas not entirely snow covered. From the National Aeronautics and Space Administration (NASA) satellite with MODIS (MODerate-resolution Imaging Spectroradiometer).
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masses include true glaciers, which are defined as moving masses of ice that are crevassed and are commonly bounded upslope by a master crevasse (bergschrund), where they have pulled away from the rocky wall of the cirque. They also include ice patches and permanent snowfields in areas largely protected from direct sunlight. The ice occurs preferentially in north-to-northeast-directed cirques, which are largely shaded, especially in winter.

From south to north, the median elevation of the ice toes decreases 500 m from 3750 m at 36.5°N to 3250 m at 38.5°N latitude, that is at 2.6 m/km of latitude (Fig. 9). The glaciers are small—most are less than 500 m long; the longest is the Palisade Glacier at 1.45 km in length. Seven exceed 800 m (Fig. 10). The elevation difference between head and toe of most glaciers is less than 200 m, and only eight exceed 400 m (Fig. 11).

Shrinkage of the glaciers has been appreciable in historic time, as shown by repeat photography. From 1900 to 2004, the area of 14 of the largest Sierra Nevada glaciers decreased 31%–78% (average 55%; Basagic and Fountain, 2011). In 1972, glacial ice and permanent snowfields covered ~50 km² (Raub et al., 2006).

Outlines of the satellite-determined present-day summer snowpack show a monthly shrinkage that eventually converges on the elevation and position of the mapped ice masses (compare Figs. 4, 5, and 8). This convergence of the area covered by the residual summer snowpack with that covered by the present-day glaciers occurs late in the season—by July in a low snow year and by September in a high snow year. It is this late summer residual snowpack that provides the nourishment that sustains the present-day glaciers and ice masses year after year. When the snowpack is totally melted earlier in the season, the glaciers recede.

Rock Glaciers

Rock glaciers are tongue-shaped masses of ice covered by bouldery rock debris that are geographically restricted to the higher parts of the southern Sierra Nevada. They occur in and near the same area as the ice masses (Fig. 12). They nestle generally in northerly facing cirques with an average flow direction of N10°E. Arcuate ridges in their lower reaches are parallel to the steep lobate toe. Ongoing movement of most rock glaciers is demonstrated by a steep toe scarp, up to 100 m high, which is mantled by large fresh angular boulders balanced at the angle of repose. Disruption of pack trails by movement of rock glaciers is occasionally reported.

Some 154 rock glaciers are depicted on geologic maps at 1:62,500 scale (Caudill, 2005). Raub et al. (2006) identified 33 rock glaciers in their aerial photograph study. Smaller rock glaciers and older inactive rock glaciers have been noted particularly in the northern part of their range (Millar and Westfall, 2008). Field study of several such features indicates that some, and probably many, consist of glacial ice under thin (~1–10 m), but continuous, cover of rockfall-generated bouldery debris (Clark et al., 1994).

Although the rock glaciers occur in about the same span of the range as ice and permanent snowfields, they are more concentrated toward

Figure 6. Width of the snowpack as measured along the 38th parallel from the beginning of March to the beginning of June during the years 2000–2011. Note that on 1 May in the average years of 2004, 2005, 2006, 2008, and 2009, the width of the snowpack at the 38th parallel is ~52 km, i.e., the same width as the lake zone. From the National Aeronautics and Space Administration (NASA) satellite with MODIS (MODerate-resolution Imaging Spectroradiometer).

Figure 7. Elevation of timberline and Sequoia groves as plotted from U.S. Geological Survey topographic maps. Trees reach to the Sierra Nevada crest south of 36.3°N and north of 38.2°N latitude.
the south (Fig. 12). On average, the elevation of the toes of exposed ice masses is distinctly higher by 200–300 m than the toes of rock glaciers (Fig. 9). This results apparently from the insulating layer of rocky rubble that protects the internal ice of the rock glaciers from the sun’s heat, as well as from melting induced by wind (Clark et al., 1994). The greater concentration of rock glaciers in the south (Fig. 12) may result from lower precipitation, which would tend to increase the ratio of cirque-wall talus to snow. The rock glaciers favor coarse-grained granitic rocks, which produce coarse blocky talus, but some occur in areas of metavolcanic rocks, as in the Ionian Basin.

The median elevations of both rock glacier toes and glacier and ice mass toes descend systematically toward the north—the rock glaciers at 3.4 m/km and the ice at 2.6 m/km of latitude (Fig. 9). In one area near 37.6°N latitude, some rock glaciers are notably lower in elevation, at 2600–2800 m (Fig. 9). These features in the Mount Morrison 15 min quadrangle appear to be older, inactive rock glaciers (Rinehart and Ross, 1964).

Rock glaciers are commonly 100–1000 m in length, with a median length of 550 m (Fig. 13). Their maximum size is similar to that of glaciers, but, surprisingly, few small ones, comparable to small glaciers and patches of ice, have been mapped. Probably, most mappers would include small rock glaciers with talus.

**PLEISTOCENE GLACIATION**

In the Pleistocene ice ages, the ice covered an area of ~20,000 km², i.e., three orders of magnitude greater than at present. Roughly a third of the range was mantled with ice. It covered broad areas above 1500 m in elevation in the north and above 2500 m in the south. Several of the large western trunk glaciers descended below 1000 m. The dating of Pleistocene glacial stages in the Sierra Nevada is hindered because of the overriding of earlier moraines by later glaciers, the erosion of glacial drift in steep canyons, and the limited location and age of datable adjacent volcanic rocks. The original four units defined by Blackwelder (1931)—McGee, Sherwin, Tahoe, and Tioga—are still the basis for Sierra Nevada glacial chronology, despite problems uncovered by more recent work (Gillespie et al., 1999). Recent studies have expanded these four to as many as 15 (Gillespie and Zehfuss, 2004), and future work will no doubt increase the number. The oldest relatively well-dated deposit is the Sherwin till, dated at ca. 820 ka (Clark et al., 2003). It underlies the Bishop tuff (Kistler, 1966), dated at 767 ka (Crowley et al., 2007).
Figure 10. Length of glaciers and ice masses in the Sierra Nevada (Raub et al., 2006). Selected larger glaciers are named.

Figure 11. Elevation of head and toe of 1285 ice masses in the Sierra Nevada (Raub et al., 2006). Glaciers with vertical extent up to 200 m and 400 m are indicated.
The mapping and separation of moraines on the lower west side of the range are limited. Matthes (1930) has mapped the extent of both Wisconsin and the deeply weathered pre-Wisconsin drift (El Portel) in the Yosemite region. Likewise, he has differentiated Wisconsin moraines in the San Joaquin River drainage from an older, more extensive pre-Wisconsin drift (Matthes, 1960). The Wisconsin, he suggests, includes both the Tahoe and Tioga stages of Blackwelder (1931), and the pre-Wisconsin correlates with Blackwelder’s Sherwin glaciation (Matthes, 1960, p. 48; Huber, 2007).

A geochronological study of the Tahoe and Tioga moraines in the Bishop Creek drainage of the southern Sierra Nevada, utilizing the accumulation of cosmogenic $^{36}$Cl, indicates that the moraines mapped as Tahoe stage formed at 170–130 ka, whereas those in the Tioga stage fall in four groups from 28 to 14.5 ka, and those in the Recess Peak stage formed from 13.4 to 12.0 ka (Phillips et al., 2009). A second study utilizing $^{10}$Be in samples from the same and additional moraine on the east side of the Sierra Nevada found Tahoe stage moraines at 144 ± 14 ka and Tioga stage moraines at 19 ± 1.9 ka (Rood et al., 2011).

These studies indicate that the Tahoe stage in the southern Sierra Nevada corresponds with the global marine oxygen isotope stage (MIS) 6. A time hiatus of more than 100 k.y. is indicated between the Tahoe and the Tioga glacial stages.

In contrast, cosmogenic dating of Tahoe moraines in the type locality at Lake Tahoe by the $^{10}$Be and $^{26}$Al methods indicates an age of 69.2 ± 4.8 ka, equivalent to the global marine isotope stage MIS 4 (Howle et al., 2012). Hence, moraines mapped as Tahoe in the northern Sierra Nevada are apparently only ~50 k.y. older than the Tioga. One interpretation is that in the southern Sierra Nevada, the “younger” Tahoe did not advance as far as the later Tioga, so that its moraines were obliterated by advance of the Tioga ice. In contrast, in the northern Sierra Nevada, the “younger” Tahoe advanced further than the “older” Tahoe and thus obliterated its moraines, yet these were not covered by the Tioga ice.

Limited geochronology has been done on northern west-side glaciers. A $^{10}$Be study of moraines in Bear Valley found that Tioga stage glaciation in the valley peaked 14,100 ± 1500 yr ago (James et al., 2002). In addition, they noted the presence of considerable areas of morainal deposits older than Tioga stage glaciation.

Recess Peak stage moraines occur high in the cirques just below the small present-day (Matthes) glaciers. They were initially believed to be nearly as young as the present glaciers, but
the work of Phillips et al. (2009) indicates that they were deposited between 13.4 and 12.0 ka (as compared to ca. 14.2–13.0 cal yr B.P. as determined by Clark, 1997). The fact that the high-level Recess Peak advance is only a few thousand years younger than the last of the preceding Tioga glaciations demonstrates the very rapid retreat from the much more extensive Tioga positions. It is this rapid retreat that apparently exposed most of the glacial lakes as we see them today, and it helps to account for the paucity of morainal material in this heavily glaciated sector above the Tioga moraines.

**Lakes**

The U.S. Geological Survey 1:24,000 scale quadrangle maps depict 7040 lakes in the range, which lie in a nearly continuous N30°W-trending belt that is 450 km long (390 km of N latitude) in all but the southermmost 100 km of the range (Fig. 1). Small numbers of lakes are dammed by landslides, fault scarps, or are of volcanic origin. The lake listings were edited to exclude these as well as artificial lakes dammed as reservoirs and occurring at obviously low elevation. The largest lake in the Sierra Nevada, Lake Tahoe (Fig. 1), into which several Pleistocene glaciers flowed from the west and south, is a structural depression formed by basin-range faulting and enlarged by lava-flow damming.

Some of the major glacial valleys on the west side of the range have been artificially dammed, producing reservoirs in their glaciated course. These include Salt Spring Reservoir on the North Fork of the Mokelumne River and Hetch Hetchy Reservoir on the Tuolumne River. Many small ponds constructed for logging operations or dug in pastures for cattle watering were deleted. Lakes smaller than 0.001 km² in area (~36 m in diameter) were deleted, as were other small ponds that proved to be wide spots in streams.

Most lakes are of glacial origin and were carved by ice in bedrock and impounded by bedrock sills. For example, of the 302 lakes in the Mount Whitney 15 min quadrangle (Moore, 1981), only 80, or 26%, are dammed by moraine or talus, while the others have bedrock sills. The bedrock basins are produced by the erosive action of ice and basal water, with the most vigorous excavation in those areas with preexisting closely spaced joints or fractures (Drewry, 1986). In such excavation, the ice at the base of the glacier flows uphill to erode and remove debris and in this way produces closed basins, even though the surface and bulk of the glacier are flowing downhill. These quarried lakes often occur in chains down the valley, sited on steps. Commonly, multiple such beaded lakes (termed paternoster lakes) occur in high glacial valleys below the cirques formed at the head of the glacier.

Some glacial lakes dammed by moraines form near the low-gradient, distal parts of glaciated valleys. The largest and most impressive of these are found along the east foot of the range between 37°53'N and 37°27'N lat. Here, large glaciers in the lower reaches of their canyons have dammed lakes with their terminal and recessional moraines. These include Walker Lake on Walker Creek; Grant Lake, Silver Lake, Gull Lake, and June Lake on Rush Creek; Convict Lake on Convict Creek; Hilton Lakes on Hilton Creek; and Rock Creek Lake on Rock Creek. In addition large lakes—Fallen Leaf and Cascade—are dammed by lateral and terminal moraines in the Tahoe Basin.

Other lakes form where lateral moraines of a valley glacier dam smaller tributary streams. In addition, some high-elevation lakes owe their origin to ponding within hummocky neoglacial till, and damming by rockfalls and rock glaciers from oversteepened cirque walls.

Glacial lakes in the entire range occur in the elevation range of 1500–4000 m (Figs. 1 and 2). The belt of lakes attains its greatest width of 45 km along the south half of the range, and proceeding north the width decreases at 38.5°N to 40 km, at 39°N to 35 km, and at 39.5°N to 15 km. This widening to the south parallels the greater width of high-elevation terrain (Fig. 1). The northernmost lakes occur near 39.7°N latitude, somewhat north of Sierra Butte, but the ill-defined northern extent of the Sierra Nevada makes this limit problematic. The southernmost glacial lake identified on topographic maps occurs at 36.096°N lat, 118.574°W long (Fig. 14).

In the Sequoia–Kings Canyon National Parks, lakes occur in the upper part of the Tahoe/Tioga glaciated area (Fig. 15). The lake area lies in the area of primary snow accumulation. The glaciated zone where lakes are absent occurs at lower elevations in the zone of ablation, where ice was largely restricted to the canyon-bound trunk glaciers.

Prominent embayments with no lakes on the west side of the lake belt are associated with major river canyons, where low-elevation terrain reaches deep into the range (Figs. 1 and 14). The two south-directed zones of lakes on the south end of the glaciated terrain (Figs. 14 and 15) flank the giant south-directed Kern Canyon. On the west slope, west-protruding highlands between canyons support many lakes. The three main embayments in the lakes between latitudes 36.7°N and 37°N are from canyons of the three forks of Kings River. Those from 37.2°N to 37.6°N are from forks of the San Joaquin River, and the partial gap in the crestal lakes at 37.7°N is caused by low regions of the North Fork San Joaquin River canyon leading to Mono Pass (3000 m elevation).

Proceeding north, forks of the Stanislaus and Mokelumne Rivers drain the low region near Sonora Pass (38.3°N–38.5°N lat), where a thinning of the lake zone occurs. In addition to the low elevations formed by the giant canyons of these rivers, the abundant crestal cover of Tertiary volcanic rocks may also inhibit lake development. Further north, a prominent gap in the lakes at 39.1°N–39.3°N occurs in the headwaters of the North Fork of American River, which includes Donner Pass north of Lake Tahoe. The next well-glaciated region occurs in the highlands including Bowman Mountain, Man Mountain, and Red Mountain between the South and Middle Forks of the Yuba River.

The most northerly region of glaciation (extending to 39.8°N lat) lies north of Sierra Buttes. It includes Gold Lake, Salmon Lakes, and Sardine Lakes. This lake cluster is drained on the south by the North Fork of the Yuba River and on the north by the Middle Fork of the Feather River.

The effective upper limit of the lakes (the elevation with 1% of the lakes higher) decreases northward from 3900 to 2300 m (3.4 m/km of latitude), the median elevation of the lakes decreases northward from 3500 to 2000 m (4.5 m/km of latitude), and the lower limit of the lakes (the elevation with 1% of the lakes lower) decreases northward from 2600 to 1500 m (3.1 m/km of latitude; Fig. 16). The high lakes are controlled by the elevation of the crest of the range and are about coincident with the median elevation of existing ice. The low lakes track the lower elevation of glaciers capable of eroding lake basins that have subsequently remained unfilled by sediment and till.

The downstream slope of the lower lake level (3.1 m/km of latitude) is steeper than, but similar to, that of the present-day glaciers, which decrease from 3670 m at 36.6°N latitude to 3250 m at 38.4°N latitude at 2.3 m/km of latitude (Figs. 9 and 16).

**Moraines and Glacial Canyons**

Information on glacial moraines is from 15 min geologic maps, most published in the USGS series and readily accessible on the internet (Caudill, 2005). Because the moraines are preserved differently in different areas, the maps depict various degrees of detail. Many maps separate the two major younger Pleistocene moraine groups that commonly form paired deposits, the Tahoe and Tioga of Blackwelder (1931). Others simply lump all morainal material in one unit.
In only a few areas, moraines older than those of the Tahoe stage have been mapped.

On the steep east flank of the range, the major Pleistocene trunk glaciers occur closer to the crestal ice fields now represented by the glacial lakes (Fig. 14). Because of the asymmetry of the range, the main trunk glaciers east of the crest generally extended 5–20 km from the crest, while those west of the crest extended 25–60 km (Fig. 14).

Delineation of glacial advances is difficult on the heavily forested west slope of the range, where high rainfall has led to weathering and erosion of deposits and incision of deep canyons. Vigorous stream action has commonly removed much of the morainal record associated with the major trunk glaciers. In the Merced and San Joaquin River drainages, deposits older than Tahoe/Tioga have been correlated with the Sherwin glaciation on the east slope of the range (Matthes, 1930, 1960).

The areas covered by ice of the Tioga and Tahoe stages were nearly coincident. Hence, a common estimate is made in this study of the general position of the combined Tioga-Tahoe termini of glaciers in the major canyons of the west slope. Where morainal evidence is uncertain, this was done by a study of canyon topography, with the glacier termini located at the transition in cross section from U-shaped to V-shaped and the transition in map plan from straight to crooked (Fig. 14).

Lower ice limits generally were higher at the same latitude on the east side of the range than on the west side. In the southern Sierra Nevada, eastern glacier toes were above 1900 m and descended to ~1400 m in the north. In contrast, several western glaciers descended below 1000 m (Fig. 17). Among the longest glaciers were those in the canyons of the San Joaquin, Merced, and Tuolumne Rivers, which descended below 1200 m (Figs. 14 and 17).

Moraines in east-side canyons are restricted to higher elevations (Fig. 17) because of the smaller drainage area and lower precipitation east of the crest. Near the central part of the range on the east flank, high-elevation piedmont slopes at the distal part of the glaciers permitted the ice to exit the confining canyons and spread out on low-gradient slopes, causing morainal units of different age to be widely separated and well displayed (Clark et al., 2003; Phillips et al., 2009). In addition, the arid climate and limited vegetation in these areas helped to preserve the deposits and render them suitable for detailed study.

The lowest elevation reached by the main Pleistocene west-side trunk glaciers was markedly lower than the region of extensive snow accumulation and glaciation indicated by the

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Figure 14. The distribution of 7040 glacial lakes in the Sierra Nevada primarily generated by late Pleistocene Tioga glaciation and the lower limits of glaciation on both sides of the range. The range crest is shown by a dashed black line, and major drainages are noted. The zone of snow accumulation is about coincident with the terrain covered by the glacial lakes, and the zone of ablation extends from this out to the limits of Tioga-Tahoe glaciation as defined by terminal moraines and topographic features.
Figure 15. Map showing Tahoe glaciation in Sequoia–Kings Canyon National Parks (36.25°N–37.25°N lat; Moore and Mack, 2008) with centers of ice masses and glacial lakes shown.
alpine glacial lakes. In the central high part of the range, the trunk glaciers reached more than 2000 m lower than the general lower limit of extensive lake-producing glaciation (compare Figs. 14, 15, and 17).

Equilibrium Line Altitudes

The equilibrium line altitude (ELA) in a glacier represents the averaged boundary elevation between the upper accumulation zone, where more snow and ice accumulate than melt, and the lower ablation zone, where more snow and ice melt than accumulate. Above the ELA, the ice and snow accumulates season after season, continually increases in thickness, consolidates, and begins to flow downhill. Below the ELA, the flowing ice melts faster than it is delivered. As it flows downhill into warmer sites, melting increases until all ice disappears at the glacier toe. The position of the ELA moves uphill as the average temperature increases or the snowfall decreases; it moves downhill with temperature decrease or snowfall increase.

The zone of accumulation is largely controlled by factors that deliver snow to the range and enable part of it to remain until the next season. Average annual snowfall and temperature are important, and they are controlled in part by elevation. The zone of melting is also controlled by temperature and precipitation, but topography plays an increasing role once the ice begins flowing. In the accumulation zone, snow and ice are delivered by fall from the atmosphere, but in the ablation zone, they are delivered primarily by flow on the ground surface. When the ice from the accumulation zone flows downhill, it will seek out lower terrain in the major river canyons. Here, the ice streams will merge and...
speculate that in the late Pleistocene glacial have remained large through the seasons. We been much larger at the same date and would During glacial times, the snowpack would have and shape of the area of glacial lakes (Fig. 8). age snow year) closely encloses the distribution area of the snowpack of 1 May 2006 (an aver-
snow accumulation with the area occupied by

The upper part of the Sierra Nevada glacial system was erosive over a broad highland area as the evenly distributed ice in the accumulation zone moved to lower elevation. The scattered small lake basins record this erosive action. The lower part of the glacier system, where melting was taking place (the ablation zone), was largely confined to major preexisting river canyons. Few lakes were formed or survived in this zone, probably because postglacial sedimentation dominated in the lower-graded slopes.

A comparison of the area of present-day snow accumulation with the area occupied by glacial lakes shows a marked similarity. The area of the snowpack of 1 May 2006 (an average snow year) closely encloses the distribution and shape of the area of glacial lakes (Fig. 8). During glacial times, the snowpack would have been much larger at the same date and would have remained large through the seasons. We speculate that in the late Pleistocene glacial stages, the snowpack at the end of the summer melting season in September-October would be smallest and would have been comparable in size to the average present one at the beginning of May. This would then allow the previous season’s unmelted snow to be covered by snow of the next season. The resulting buildup of snow and ice would form the accumulation zone and then flow downslope. That movement carved the lake basins in the zone of accumulation and fed trunk glaciers that moved downslope to concentrate ice largely in previous river valleys. This movement downslope into warmer zones created the ablation zone, where eventually the glacier terminated as all ice melted.

The ELA is commonly approximated by two methods using empirical ratios derived from modern bare ice glaciers (Meierding, 1982). The first is to measure the snow accumulation area relative to the total area of the glacier. The altitude of the ELA is taken where the ratio of the accumulation area to the total glacier area (AAR) is empirically established, e.g., 0.65. The second method assumes a ratio of elevation of the equilibrium line relative to the total range in elevation of the glacier from head to toe (Toe-to-
headwall altitude ratio or THAR), e.g., 0.4–0.5. Series of models of the movement of ice during the last glacial advance in the Kings river drainage (Kessler et al., 2006) were performed by making assumptions on the orographically influenced glacier mass balance based on measured functions on modern glaciers in western North America (Meier et al., 1971; Mayo, 1984). Models were tested to find the ELA that produced the best approximation of the limit of glacial ice in the main Kings Canyon glacial system. The ELA that most closely approximated the system was at an elevation of 3170 m in the drainage of the South Fork of the Kings River (36.9°N lat). This places the paleo-ELAs somewhat below the median elevation of the lakes (Fig. 17).

An estimation of the ELAs for the main drainages of both the west and east slopes of the range was made by using the toe-to-headwall-average-ratio (THAR) method. For each major drainage, the termini elevation of late Pleistocene glaciers (primarily Tioga stage; Fig. 17) was compared with the headwall elevation of the same glacial system. The average of these two elevations, when assuming a THAR ratio of 0.5, became the ELA estimate. The results (Fig. 18) indicate an average west slope ELA (based on 22 glacier measurements) that is nearly parallel to, and 400–600 m below, the average east slope ELA (based on 32 glacier measurements). Moreover, the elevation trend with latitude of the west side ELAs is parallel with, and only 200–300 m above, that of the elevation of the lowest west side lakes (Fig. 18).

This correspondence of the rangewide west slope ELA calculated from the distribution of the late Pleistocene ice with the lower limit of the lake zone suggests that the lake zone can be a rough gauge of the position of the accumulation zone. Therefore, the generalized position of the late Pleistocene ELAs occurs close to, but somewhat above, the lower limit of the glacial lakes.

CONCLUSIONS

The 600-km-long Sierra Nevada underwent extensive Pleistocene glaciation in its northern 500 km. In this work, new and existing data are evaluated to examine the rangewide weather and the distribution of the present-day snow-pack, glacial lakes, existing glaciers and rock glaciers, and Pleistocene glaciers as determined by mapped moraines and topographic evidence. Presently ~1700 small glaciers and ice masses at 3000–4000 m elevation occur along 240 km of the range crest and cover ~50 km². These glaciers are generally less than 500 m long, and the median elevation of the lower limit of the glaciers descends north at 2.4 m/km. Repeat photography shows a historical shrinkage, with 14 of the largest glaciers having lost on average about half their area from 1900 to 2004.

About 154 rock glaciers show a similar distribution as the present-day glaciers but average 150–200 m lower. They apparently owe their lower elevation to the insulation and wind protection provided by a mantle of bouldery debris on ice.

The base of the present-day snowpack, the snowline, generally decreases in elevation