Testing a New Climatic Interpretation for the Tahoe Glaciation

Alan R. Gillespie
Quaternary Research Center
University of Washington
Seattle, Washington 98195

Abstract. The age of the Tahoe glaciation is uncertain; estimates range from ~200 to ~50 ka. Radiometric dates and pluvial records suggest that at least some glaciers of the Tahoe stade occurred after ~90 ka. Within the interval 90 to 50 ka are two different climatic periods that may have caused the glaciation: marine oxygen isotope Stage 4 (~75-65 ka), with winter Pacific storms, and a hypothesized monsoonal episode around 85 ka, with summer precipitation from the south. These different climatic regimes should impress characteristic signatures on north-south trends of glacial equilibrium-line altitudes (ELAs). ELAs for four Pleistocene and three Holocene Sierran glaciations, together spanning the last ~200 ka, were calculated by the accumulation-area ratio technique, using published maps of glacial deposits on the east side of the Sierra Nevada. ELAs for the Tioga glaciation were subparallel to the modern snowline. ELAs for the Tahoe glaciation were significantly discordant with the normal pattern, but this discordance was eliminated if the Tahoe glaciation was subdivided into early and late advances. The late Tahoe glaciation occurred during a climatic period much like the Tioga glaciation, dominated by westerlies and Pacific storms. The most likely time of glaciation was during the early stages of the latest continental glaciation, perhaps in marine oxygen isotope Stage 4. Large mountain glaciers early in the last glacial cycle suggest that

1) the marine record is not an adequate proxy for continental climate at temperate latitudes, and 2) the severity of climatic change at the end of the interglacial may have been greater than generally appreciated. Current data are insufficient to characterize the early Tahoe glaciation, but it and the Mono Basin glaciation may be the same.

INTRODUCTION

The Sierra Nevada Range of eastern California was repeatedly glaciated during the Quaternary Period. Seven Pleistocene and three Holocene glaciations have been named (see Fullerton [1986] for a recent review). Of the Pleistocene glaciations, only the four youngest are included in this study: Tioga, Tenaya, Tahoe, and Mono Basin, in order of increasing age. Blackwelder [1931] considered the Tioga and Tahoe both to have occurred during the continental Wisconsin glaciation (the Tenaya and Mono Basin glaciations were recognized later by Sharp and Birman [1963]). Previous workers have commonly distinguished early and late Tioga glaciations, and there is evidence (below) that the Tahoe glaciation may be similarly subdivided. The three Holocene glaciations (neoglaciations) are Hilgard, Recess Peak, and Matthes (decreasing age).

The Tahoe glaciation has been controversial because its age has been difficult to determine (estimates range from ~200 to ~50 ka) and because its extensive moraines have been difficult to reconcile with existing models of late Pleistocene climate [Gillespie, 1984]. At least part of the difficulty appears to be inconsistent application of relative weathering criteria among different researchers. As presently identified, some Tahoe glaciers appear to be >130 ka old and may have advanced during the penultimate continental glaciation (e.g., Burke and Birkeland [1979]; Gillespie [1982]; Porter et al. [1983]; Dorn et al. [1987]); others are younger than 120 ka but significantly older than the 25 to 11 ka Tioga glaciers [Gillespie, 1982; Gillespie et al., 1984]. These "late Tahoe" glaciers could not have coincided with the Laurentide ice sheet at its maximum extent, which occurred during marine oxygen isotope Stages 2 and 6 (e.g., Shakleton and Opdyke [1973]). Relative-weathering criteria defining the stratigraphic Tahoe stade appear to be broad enough to encompass at least two different ages of glaciation [Gillespie, 1984]; at Bloody Canyon, Mono County, a disputed "Tahoe" moraine [Sharp and

Birman, 1963; Burke and Birkeland, 1979] on further study appeared to be a composite of two distinct tills [Mathieson, 1981; Gillespie, 1982].

Smith [1979, 1984] suggested that the Sierran glaciations coincided with high stands of Searles Lake. He has shown that the lake level was high after ~90 ka, following a dry period earlier in Stage 5. Thus Searles Lake rose well before the beginning of the latest continental glaciation at ~75 ka and well after a brief glacial-like period of insolation around 115 ka during interglacial Stage 5. If the younger Tahoe glaciation coincided with an interglacial high stand of Searles Lake, perhaps it occurred at ~85 ka, a time when insolation patterns suggest that the southern Great Basin climate may have been characterized by summer storms from the Gulf of California. The similar early Holocene monsoonal climate (e.g., COHMAP members [1988]) evidently produced neoglaciation in the mountains of New Mexico [Wesling, 1988], and perhaps the Hilgard glaciation in the Sierra Nevada as well, which may have been coeval [Mezger, 1986].

Thus, the younger Tahoe glaciation may have occurred at ~70 ka, during a climatic period similar to the later glacial maximum, dominated by Pacific storms and westerlies; or it may have occurred at ~85 ka, during a time characterized by summer precipitation and cloudiness. It is unlikely to have occurred after Stage 4, because weathering differences compared to Tioga tills are so profound.

The distribution of coeval glaciers must reflect the climatic conditions under which they formed. For instance, under "normal" glacial conditions, precipitation in the Sierra Nevada originated mainly from winter storms from the Pacific Ocean, which weakened to the south. This pattern, basically an intensification of the modern one, led to geographic gradients in the average firn line or glacial equilibrium-line altitudes (ELAs): rising to the south and to the east (e.g., Porter et al. [1983]). Wesling [1988] found in New Mexico that these gradients reversed during the "monsoonal" neoglaciation. Thus, the hypothesized "monsoonal" Tahoe glaciation might differ from the Tioga glaciation (and from the hypothesized "westerly" Tahoe glaciation) in that the snowline would be lower than normal in the southern Sierras. The ELAs, normally subparallel to the modern snowline which rises southward at ~2 m/km (~11 ft/mi) [Burbank, in prep.], would have an abnormally low slope when plotted against latitude. Figure 1 depicts

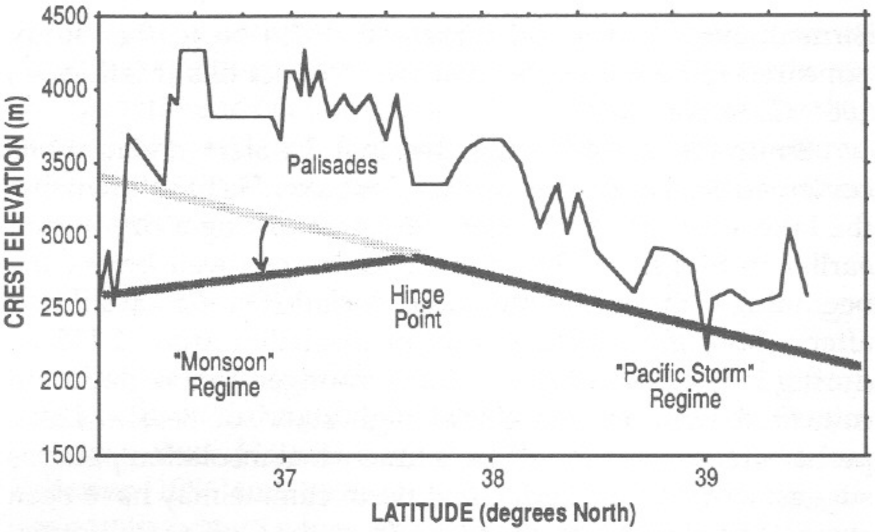

Figure 1. Schematic representation of ELA latitude gradients for the Sierra Nevada. ELAs rise southwards, parallel to the modern snowline, for glaciations caused by winter precipitation from Pacific storms. Summer precipitation and cloudiness from the south would depress ELAs in the southern Sierra Nevada. To the north, this effect would be negligible, and the ELAs would be controlled by winter storms.

schematically the possible effects of a southern source of moisture and clouds on ELAs up and down the crest of the Sierra Nevada, assuming that the summer precipitation occurred during an interval of enhanced winter storms as well. Because the ELA of a glaciation can be reconstructed from the distribution of moraines and cirques, it is possible to test the above hypotheses for the origin of the Tahoe glaciation. By determining which climatic period produced the Tahoe glaciation, it is also possible to improve current estimates of its age.

METHODS

ELAs for five Pleistocene and three Holocene Sierran glaciations, together spanning the last ~200 ka, were calculated by the accumulation-area ratio (AAR) technique, using published maps of glacial deposits on the east side of the Sierra Nevada [Bateman, 1965; Birman, 1964; Birkeland, 1963; Bursik, 1989; Clark, 1967; Fleisher, 1967; Gillespie, 1982; Janda, 1966; Mezger, 1986; Sharp, 1969, 1972]. Moraines designated "pre-Tahoe" were here considered to be Mono Basin. ELAs were also calculated for active glaciers. Figure 2 shows the areal distribution of the glaciated valleys for which ELAs were calculated in this study. This study includes only valleys east of the crest of the Sierra Nevada to minimize effects of the Sierra rain shadow. Although focus was on the Tahoe glaciation, ELAs were calculated for the entire sequence of late Pleistocene glaciations in order to provide a more complete context for interpretation. Holocene glaciations were included because one may have occurred during a monsoonal period.

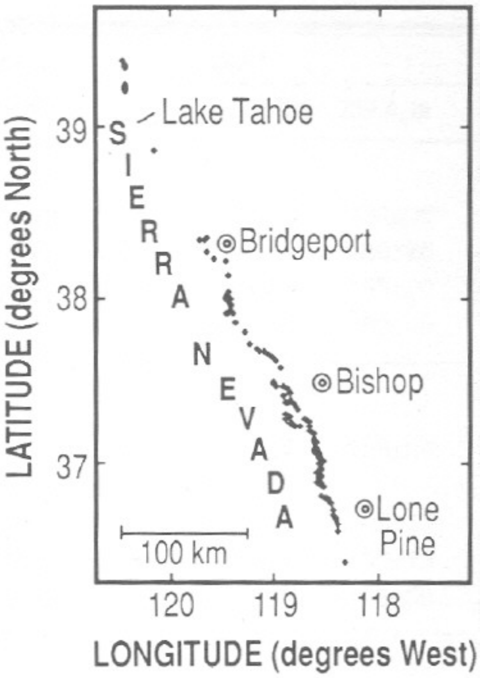

Figure 2. The distribution of glaciated valleys (dots) for which ELAs were calculated.

The AAR technique (e.g., Meierding [1982]) is based on the observation that, for active temperate alpine glaciers, the firn line divides the upper two-thirds of the glaciated valley (accumulation area) from the lower third (ablation area). Thus, the ELAs for vanished glaciers may be calculated from a map if the moraines and trim lines are known. The appropriate critical area ratio varies with steepness of the valley, but Meierding [1982] used a value of 0.65 for Pleistocene glaciers in the Rockies, and that value was adopted by Mezger [1986] for the Cottonwood Basin in the southern Sierra Nevada. It is used in the present study also. Obviously, ELAs for transfluent glaciers and outlet glaciers fed by ice caps will be be underestimated by this technique; seriously affected valleys were therefore disregarded in this study.

In principle, the highest occurrence of lateral moraines should also be near the firn line. Where possible, the ELA estimated this way was compared to the ELA calculated by the AAR technique. Agreement was within the ~100 m (~328 ft) uncertainty estimated by Mezger [1986] for the AAR technique.

ELAs estimated by the AAR technique appear to be sensitive to differences in terrain in addition to the average slope of the glacier. For example, shallowly glaciated benches below cirques are common in the Sierra Nevada and increase the accumulation area disproportionately to the increase in ice volume above the firn line; thus the ELA is underestimated with respect to an adjacent valley of simpler morphology. However, it is likely that errors of this sort will repeat for successive glaciations in the same valley. Because the main goal of this study was to compare regional ELA trends for different glaciations, the differences in ELAs for each valley (ΔELA) were also calculated. ΔELAs were always referred to the early Tioga glaciation. Calculating ΔELAs suppressed "noise" and clutter in the latitude gradients.

AARs were found by planimetry from published topographic data. Where feasible, this was done by computer, using USGS 7.5-minute and 1-degree digital topographic images. The 1-degree maps were used for larger glacier systems. The procedure was to outline the glacier/rock contact (trim line and lateral moraine crests) interactively. The ice surface within this polygon must be estimated by interpolation from the altitudes along the contact. The ELA was then found from the histogram of the altitudes for the "pixels" within the interpolated polygon. This approach is attractive mainly in the ease of calculation, but it suffers from the same drawbacks as analog planimetry: the ice surface must be assumed, and the glacier margins must be delineated correctly, by hand. Where digital topographic data were not available, published USGS 15-minute contour maps were analyzed. The two approaches agreed within the uncertainty estimated by Mezger [1986].

Small glaciers appear to be particularly sensitive to local influences, especially of topography and weather patterns. For this reason, Burbank [in prep.] tabulated ELAs only for valleys meeting a set of limiting criteria: for example, northeast aspect, minimum size, etc. Primarily because some of these effects appear to be less important for the large Pleistocene glaciers, no editing according to the criteria of Burbank was performed for this study.

ELAs were tabulated according to glaciation and geographic coordinates found for the nominal center of the accumulation area. Classical regression lines were fitted to the ELAs for each glaciation distributed by latitude. Slopes and regression coefficients *(r)* for these lines were tabulated along with the ELAs. Finally, ΔELAs were calculated by subtracting Tahoe (or other) ELAs from Tioga ELAs, for valleys in which deposits of both ages had been identified. All these results were displayed graphically against a topographic profile of the crest of the Sierra Nevada.

RESULTS

ELAs for paleoglaciers from more than 70 valleys of the eastern Sierra Nevada were calculated (Figure 2). Statistical summaries of ELAs for active glaciers and the seven glaciations are given in Table 1. Early Tioga ELAs for individual valleys are reported in an altitude versus latitude plot, shown in Figure 3. Individual ΔELAs are shown in Figures 4 and 5.

Table 1 shows that paleo-ELAs are depressed below the modern values by ~900 to 240 m (~2,953 to 787 ft), with each successively older glaciation depressed a greater amount. Depression from the "true" modern regional snowline would be 100 to 200 m (328 to 656 ft) greater [Meierding, 1982]. Thus, the late Pleistocene ELAs were depressed at least 700 m (2,297 ft). It is noteworthy that, at one site, both Matthes and Recess

Table 1

Glaciation[1]	ELA				Δ ELA[2]			
	N	*r*	slope, m/km	ELA, m	*N*	*r*	slope, m/km	ΔELA, m
Holocene Period								
Active	64	0.45	−1.8±0.5	3740±30	59	0.06	0.3±0.7	720±25
Matthes	38	0.57	−2.4±0.6	3500±30	38	0.31	1.2±0.6	480±25
Recess Peak	30	0.88	−3.1±0.3	3360±30	30	0.19	0.3±0.3	335±20
Hilgard	18	0.89	−3.9±0.5	3190±30	30	0.19	−0.3±0.3	160±40
Late Pleistocene Period								
Tioga, late	11	0.64	−2.2±1.0	3120±40	11	0.01	−0.0±0.7	125±30
Tioga, early	70	0.90	−3.1±0.2	3040±20	—	—	—	0
Tenaya	17	0.69	−2.3±0.7	2975±65	17	0.09	−0.1±0.3	−45±15
Tahoe	40	0.89	−3.0±0.3	2910±40	39	0.46	0.4±0.1	−155±15
Mono Basin	9	0.82	−3.4±1.1	2875±75	9	0.22	−0.4±1.6	−195±50
Subdivided Tahoe glaciation								
Tahoe, late	24	0.97	−3.3±0.2	2940±30	24	0.52	−0.2±0.1	−95±10
Tahoe, early	12	0.19	−0.6±1.0	2800±50	12	0.01	0.1±0.6	−230±25

1. In order of increasing age. See Gillespie [1982] and Fullerton [1986] for reviews.
2. $\Delta ELA = ELA - ELA_{early\ Tioga}$

*Table 1. Statistical summary of eastern Sierra Nevada equilibrium-line altitudes. **N**—Sample size (may be different for ELA and ΔELA). **r**—linear regression coefficient Slopes are negative if ELA rises southward. Values of ELA and ΔELA are calculated for 37° N. Modern ELA at 37° N is near 3,860 m at 37° N and rises ~1.8 m/km southward according to Burbank [1990], based on selected glaciers. The "true" modern regional snowline may be 100 to 200 m higher [Meierding, 1982]. The altitude of the Sierra Nevada crest at 37° N is ~3,900 m. Reported uncertainties are ±1 σ (standard deviation). They do not include analytic precision for each calculated ELA.*

Peak ELAs were ~390 m (~1,280 ft) higher than in neighboring valleys. This anomalous site is the type of area defined by Birman [1964] for the Recess Peak glaciation, in the Second Recess of Mono Creek.

Lines regressed to the ELA versus latitude distributions explained from 20 to 79 percent of the variance of the data. The worst fits were for the active and Matthes glaciers, which were restricted to cirques. Regression to the late Tioga and Tenaya ELAs accounted for less than half the variance also. Burbank [in prep.] reported that regression to ELAs for selected active glaciers explained 50 percent of the variance, a significant improvement over 20 percent for the unedited glaciers.

The regressed lines for all the glaciations had negative slopes (ELAs rose southward), and all the 2-σ error bars overlapped the early Tioga ELA except for the active glaciers. The early Tioga data gave the most precise slope, which at –3.1 m/km (–16 ft/mi) was steeper than the modern gradient of –1.8 m/km (–10 ft/mi) found by Burbank [in prep.] for the southern half of the present study. The uncertainty in the slope (Table 1)

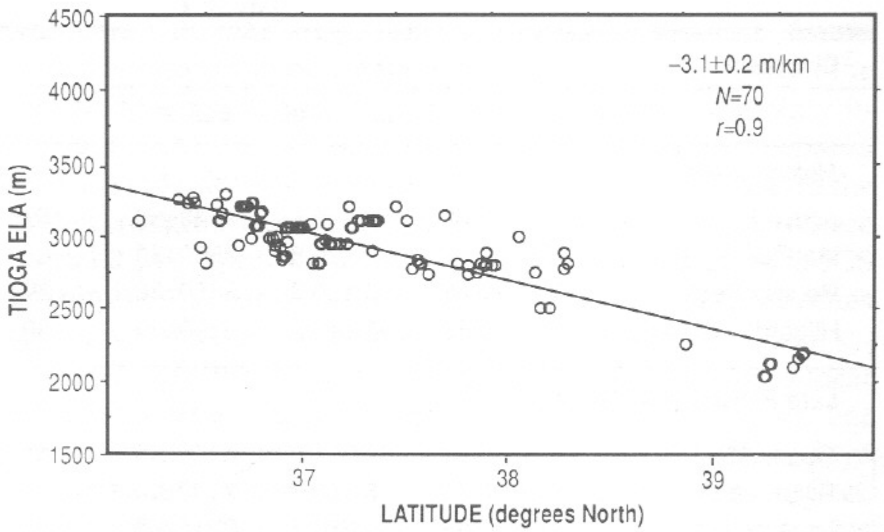

Figure 3. ELA and regression line for the early Tioga glaciation versus latitude. Slope of the line, number of points, and regression coefficient are shown in upper right.

was twice that found in a Monte Carlo simulation using the precision estimates of Mezger [1986].

The Hilgard ELAs alone rose southward faster than Tioga ELAs, at the 1-σ level. Of particular interest to this study, is that the trends for the Tahoe and early Tioga glaciations agreed within 1 σ.

Tioga and Tahoe tills were the most widely recognized and mapped. In the sources referred to in this study, there seemed to be confidence in identifying Tioga moraines in particular, and early Tioga ELAs were therefore chosen as a baseline for this study. Figure 3 shows that even these data scattered ±250 m (±820 ft) relative to the regressed line. There was also a tendency for the early Tioga ELAs to plot above the line from ~37 to 38° N and below it elsewhere. The data are fit just as well by a convex parabola as by a line. If only the data south of 38° N are considered, the Tioga ELAs appear to rise only 2.2±0.3 m/km (12±2 ft/mi) southward, significantly less than the modern trend. This slope is close to the estimate of Burbank [in prep.] for the same geographic region. If the tendency to bow upwards near 37 to 38° N results from some effect of terrain, it should not be detectable in ΔELA data provided that the climatic regime did not vary.

Table 1 shows that ΔELA trends for all the glaciations except Tahoe were the same as Tioga, at the 2-σ level. The ELA slope for the Hilgard glaciation was brought within 1 σ of Tioga by the differencing. The slope for Tahoe was positive; i.e., the separation between Tahoe and Tioga ELAs was less at 39° than at 37° N (50 and 150 m [164 and 492 ft], respectively). It differed from Tioga at the 4-σ level (<1percent chance that this difference was not significant).

Because ELAs for the larger glaciations covaried strongly (r>0.8), ΔELA slopes clustered around zero, and uncertainties in regression coefficients were reduced by the differencing. The precision of the ELA displacements was likewise generally improved, although the values themselves were stable. In contrast, values of r were lower: regression explained less of the variance of the differences than of the ELA data themselves, presumably because differencing preferentially retained the part of the data that was poorly correlated with the Tioga ELAs and hence with latitude. Thus, the degree of discordance or concordance of different ELA trends can better be determined using ΔELAs than ELAs. The ΔELA data for are displayed in Figure 4.

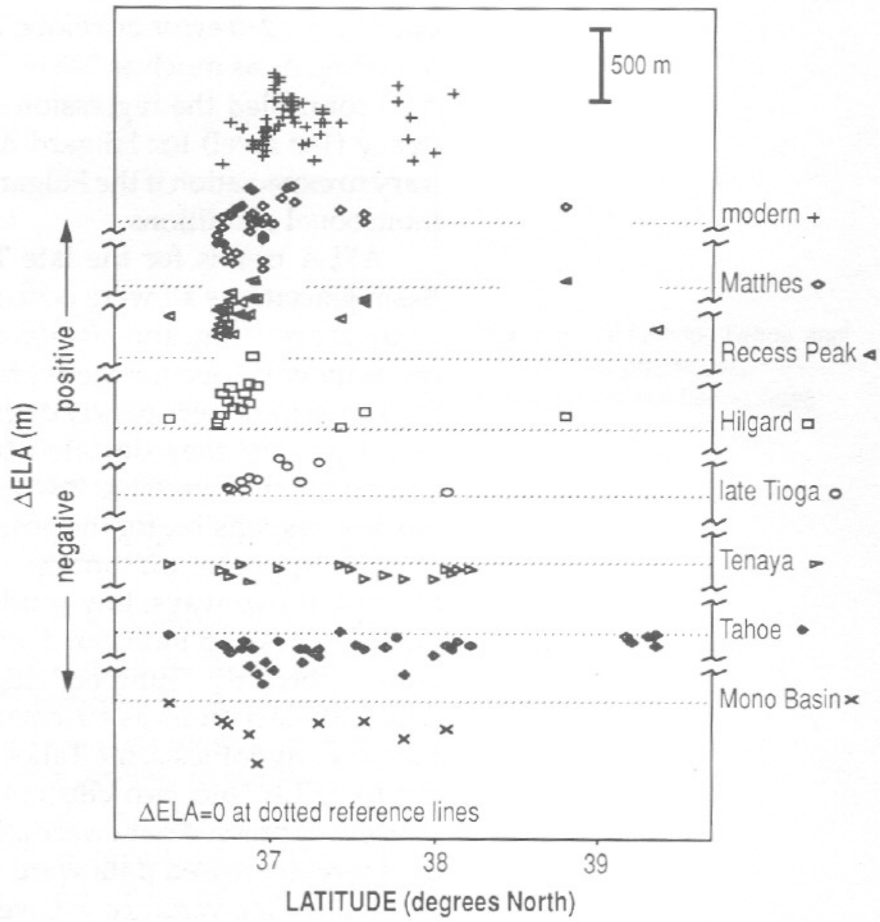

Figure 4. ΔELA versus latitude for Sierra Nevada glaciations. Ordinate is offset 500 m (1,640 ft) for each glaciation for clarity. Dotted reference lines for each glaciation are for ΔELA = 0. Data plotting below a reference line indicate lower ELAs than for the early Tioga glaciation.

ΔELA trends for the modern and Holocene cirque glaciations showed considerable disorganization, although less than the ELA data themselves. For example, the anomalously high Matthes ELAs in the Second Recess were matched by anomalously high Tioga ELAs: the ΔELA value was not deviant (this was also the case for the Recess Peak data). Active ELAs deviated ±300 m (±984 ft) from the regressed line. Generally, from 37° to 38° N the deviations were positive; to the south they were negative. The transition point does not occur at a boundary between mappers, but it does occur near the south end of the 4,000 m (13,100 ft) high Palisades region of the crest of the Sierra Nevada (see Figure 1). Matthes ΔELAs dropped weakly to the south, but this trend was controlled by the sparse data from the northern Sierra Nevada.

Recess Peak data scattered widely and showed a weak tendency to drop southwards also. Twenty points plotted outside the 2-σ error envelope containing the regression line, deviating by as much as 225 m (738 ft). Again, sparse northern data controlled the regression slope. There was a weak tendency (1-σ level) for Hilgard ΔELAs to rise southward, contrary to expectation if the Hilgard glaciers were created during monsoonal conditions.

ΔELA trends for the late Tioga, the Tenaya, and Mono Basin glaciations all were concordant with early Tioga within analytic precision, and few data points lay outside the 2-σ uncertainty envelope for the regressed lines (not shown). Only the Tahoe trend was poorly defined: 20 points were outside the envelope, and they deviated by as much as 200 m (656 ft). The points that were too low were concentrated in the south; this was responsible for the positive slope of the regressed line.

The apparent discordance of the Tahoe data can be explained in two ways: low southern ELAs are consistent with the hypothesized monsoonal origin, or Tahoe data may have been incorrectly "lumped" together, actually belonging to different glaciations as suspected at Bloody Canyon. To test this latter hypothesis, the Tahoe data were subdivided according to ΔELA into two clusters by the –120-m (–394-ft) line. The less depressed data were assigned to a "late Tahoe" stade; the more depressed data were "early Tahoe." Table 1 shows that late Tahoe variance was very well modeled by a straight line having the same slope as both the early Tioga and modern trends. In contrast, the early Tahoe was poorly correlated with latitude, and the slope of the regressed line was significantly different. Too few data, distributed over too small a range of latitudes, comprise the early Tahoe cluster to allow good statistical characterization. Figure 5 depicts the ΔELA data for the subdivided Tahoe and the Mono Basin glaciations. It is noteworthy that the early Tahoe data overlie the Mono Basin cluster.

DISCUSSION

No significant changes in trends were detected within the Holocene ΔELA data. In part, this was no doubt due to the high degree of scatter of the data, especially for the cirque glaciations, and to the paucity of data from the northern Sierra Nevada. The high degree of scatter for the cirque glaciations may have resulted from idiosyncratic effects of terrain, as predicted by Burbank [in prep.]. Nevertheless, ELA slopes for

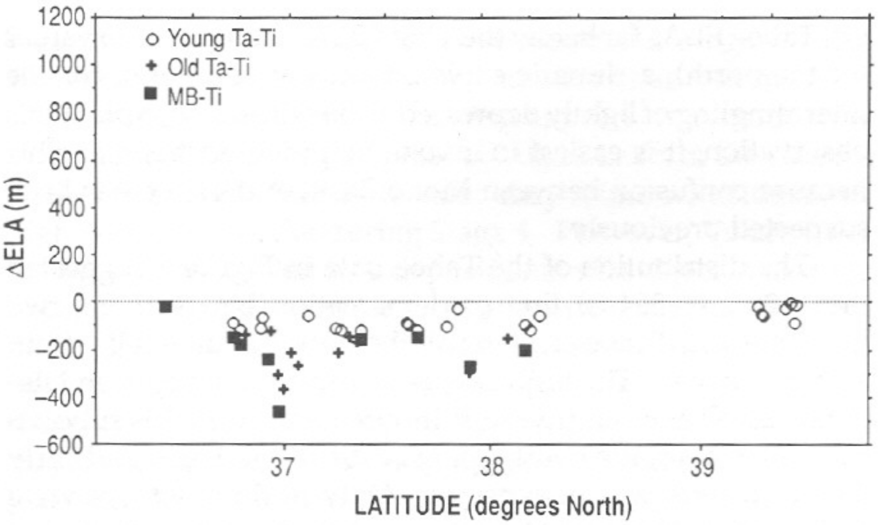

Figure 5. ΔELAs for Tahoe and Mono Basin glaciations. Ordinate has not been offset.

the cirque glaciations agreed with Burbank's better constrained results.

The Recess Peak and Hilgard ELAs appeared to rise southward more rapidly than the ELAs of Burbank [in prep.] and were subparallel to the late Pleistocene ELAs found in the present study. It may be that this disagreement is only apparent, resulting from the lack of active glaciers in the lower Sierra Nevada, north of ~38.5° N. In this interpretation the more negative slopes of the older ELAs result from inclusion of northern data. These northern ELAs are lower than predicted from the trend for the southern Sierra Nevada. This interpretation is consistent with the generalized regional patterns of cirque elevations for the western United States reported by Porter et al. [1983], but this cannot be demonstrated from the ELAs presented here. None of these data require a different climatic regime from the one during the Tioga glaciation: they can be explained by intensification of Pacific storms alone. The Tioga data permit, but do not require, differential lowering of northern ELAs. More data from the northern Sierra Nevada are required to clarify the pattern.

It is important to note that the possibility of a curved trend of Tioga ELAs requires a refined understanding of the model presented schematically in Figure 1, in which the "normal" ELAs rise linearly to the south. It is the increased *relative* depression of the southern Tahoe ELAs that would indicate an intensified monsoon, not the curvature itself.

The only Pleistocene data that appeared to deviate significantly from the Tioga data were those for the Tahoe glaciation. Although the monsoon could explain the depression of southern Tahoe ELAs far below the Tioga ELAs (compared to values for the north), a climatic explanation cannot account for the intermingling of lightly depressed Tahoe ELAs. To explain this observation, it is easiest to invoke mapping errors, plausible because confusion between Mono Basin and Tahoe has been suspected previously.

The distribution of the Tahoe data in Figure 4 suggested the –120-m (–394-ft) line of demarcation between the two hypothesized Tahoe clusters: all the northern data fall within a single cluster. The high degree of correlation between late-Tahoe ELAs and latitudes results circularly from this subdivision. However, the overlapping of the Mono Basin and early Tahoe clusters was unexpected. Only in three valleys were both early Tahoe and Mono Basin deposits mapped: on one hand this precludes direct comparison; on the other it is to be expected if the glaciations were really the same (within current limits of detectability). The three valleys containing both tills are expainable if there are multiple, currently unrecognized, Stage-6 glaciations.

The ELA for the "Tahoe" Hogsback moraine of Sawmill Canyon, Inyo County, dated by Gillespie et al. [1984] at <119±7 (2σ) ka, plots within the late Tahoe cluster in Figure 5. The Stage-6 Tahoe dates (on rock varnish; Dorn et al. [1987]) were for a glacier plotting on the boundary between clusters. Thus, the available data suggest that the late Tahoe cluster may have been deposited during oxygen stages 4 or 5 (stages 2 or 3 appear unlikely because of the profound weathering differences between Tahoe and Tioga tills). Since the late Tahoe data do not show the expected imprint of monsoonal conditions, oxygen Stage 4 may be the best age for the glaciation, provided that the Tahoe glaciation is correlated to some extent with continental glaciations, as seems to be true for the Tioga glaciation. The early Tahoe glaciers may well have occurred during Stage 6.

Because the Hilgard ELAs, from a period when monsoonal conditions are thought to have prevailed, show no "monsoonal" imprint either, it may appear that ELA trends are insensitive to climatic regimes. In this case it wouldn't be possible to refine the date of the Tahoe glaciation using ELA trends. This argument is weak because it contradicts the findings of Wesling [1988]. Perhaps the apparent insensitivity is best attributed to excessive scatter in the data and to insufficient regional extent of mapped Hilgard till. Alternatively, the

Hilgard neoglaciation may not have occurred during the latest Pleistocene/Holocene monsoonal conditions as indicated.

A large Stage-4 Tahoe glaciation appears to contradict the findings of Atwater et al. [1986] at Tulare Lake, in the Central Valley west of the Sierra Nevada: they found no evidence of high water in the lake during Stage 4. However, Tulare Lake levels were controlled by a depositional sill subject to tectonic or other disturbance, and the climatic inferences from the lake levels are thus not unambiguous. Elsewhere in the Central Valley, evidence assembled by Marchand and Allwardt [1981] showing extensive Stage-4 alluviation indicates there was a major depositional pulse then, suggesting contemporaneous glaciation in the mountains (e.g., Pierce and Scott [1982]).

The Sierra Nevada is a tectonically active range that has been rising throughout the Quaternary Epoch (e.g., Huber [1981]). Relative late Pleistocene vertical uplift rates between the Sierra Nevada and the Owens Valley appear to be between 0.1 and 0.5 mm/year (0.004-0.020 in/year) [Gillespie, 1982]. Thus, during the ~65 ka since the late Tahoe glaciation, the Sierra Nevada rose ~30 m (~98 ft) or less. This small amount is below the noise level in the present study.

CONCLUSIONS

Glacial deposits attributed to the Tahoe stade evidently were deposited during at least two different glaciations. ELAs for the early Tahoe and Mono basin glaciations were similar, suggesting that the two names refer to the same event. Burke and Birkeland [1979] showed that Tahoe and Mono Basin tills at Bloody Canyon had similar weathering and soil development, supporting this conclusion. Radiometric and rock-varnish dates from the Sierra Nevada and correlation to dated Bull Lake till in the Rocky Mountains [Pierce et al., 1976] indicate that the early Tahoe-Mono basin glaciation could have occurred during Stage 6, during the penultimate (Illinoian) ice age.

ELAs for the late Tahoe glaciation were ~100 m (~328 ft) lower than for the Tioga glaciation, which occurred during the maximum advance of the Laurentide ice sheet. It appears that the late Tahoe glaciation did not occur during the hypothesized "monsoonal" period at ~85 ka, but rather during a climatic period dominated by winter storms similar to the Tioga glaciation, but more intense. Radiometric dating, Searles Lake levels, and climatic inference from the southward rise of the ELAs implicate the early stages of the latest continental (Wisconsin) glaciation, during marine oxygen isotope Stage 4 (~65-75 ka) as the most likely time of the late Tahoe glaciation. If correct, this hypothesis requires that early in the latest ice age, mountain glaciation was more extensive than later, when the continental ice caps achieved their greatest extent. This incongruity points out the differences between the marine record of global climate and local effects on the continents at the temperate latitudes where most human societies are based. The marine record by itself is not an adequate proxy for continental climate at temperate latitudes. The onset of major climatic changes at the beginning of an ice age may be more drastic than previously appreciated, and the effects may be focused where they will have the greatest direct impact on mankind.

Acknowledgements. Aspects of the research were supported by the Land Processes Branch of NASA. Discussions in the field with Bud Burke and Malcolm Clark were especially helpful.

REFERENCES

Atwater, B. F., D. P. Adam, J. P. Bradbury, R. M. Forester, R. K. Mark, W. R. Lettis, G. R. Fisher, K. W. Gobalet, and S. W. Robinson, A fan dam for Tulare Lake, California, and implications for the Wisconsin glacial history of the Sierra Nevada, *Geol. Soc. Am. Bull., 97*, 97-109, 1986.

Bateman, P. C., Geology and tungsten mineralization of the Bishop District, California, *U.S. Geol. Sur. Prof. Pap., 470*, 208 pp., 1965.

Blackwelder, E., Pleistocene glaciation in the Sierra Nevada and Basin ranges, *Geol. Soc. Am. Bull.,* 42, 865-922, 1931.

Birkeland, P. W., Pleistocene glaciation of the northern Sierra Nevada, north of Lake Tahoe, California, *J. Geol., 72*, 810-825, 1963.

Birman, J. H., Glacial geology across the crest of the Sierra Nevada, *Geol. Soc. Am. Spec. Pap. 75*, 80 pp., 1964.

Burbank, D., Late Quaternary snowline reconstructions for the southern and central Sierra Nevada, California: Re-assessment of the Recess Peak glaciation, in prep.

Burke, R. M. and P. W. Birkeland, Reevaluation of multiparameter relative dating techniques and their application to the glacial sequence along the eastern escarpment of the Sierra Nevada, California, *Quat. Res., 11*, 21-51, 1979.

Bursik, M. I., Late Quaternary volcano-tectonic evolution of the Mono Basin, eastern California, Ph.D. diss., 270 pp., Calif. Inst. Tech., Pasadena, California, 1989.

Clark, M. M., Pleistocene glaciation of the drainage of the West Walker River, Sierra Nevada, California, Ph.D. diss., 170 pp., Stanford University, University Microfilms Inc., No. 68-6401, Ann Arbor, Michigan, 1967.

COHMAP members, Climatic changes of the last 18,000 years: Observations and model simulations, *Science, 241*, 1043-1052, 1988.

Dorn, R. I., B. D. Turrin, A. J. T. Jull, T. W. Linick, and D. J. Donahue, Radiocarbon and cation-ratio ages for rock varnish on Tioga and Tahoe morainal boulders of Pine Creek, eastern Sierra Nevada, California, and their climatic implications, *Quat. Res., 28*, 38-49, 1987.

Fleisher, P. J., Glacial geology of the Big Pine drainage, Sierra Nevada, California, Ph.D. diss., 126 pp., Washington State Univ., University Microfilms Inc., No. 68-695, Ann Arbor, Michigan, 1967.

Fullerton, D. S., Chronology and correlation of glacial deposits in the Sierra Nevada, California, in *Quaternary Glaciations in the Northern Hemisphere*, edited by V. Sibrava, D. Q. Bowen, and G. M. Richmond, pp. 197-200, *Quat. Sci. Rev., 5*, 1986.

Gillespie, A. R., Quaternary glaciation and tectonism in the southeastern Sierra Nevada, Inyo County, California, Ph.D. diss, 695 pp., Calif. Inst. Tech., Pasadena, 1982.

Gillespie, A. R., J. C. Huneke, and G. J. Wasserburg, Eruption age of a ~100,000-year-old basalt from 40Ar-39Ar analysis of partially degassed xenoliths, *J. Geophys. Res., 89*, 1033-1048, 1984.

Gillespie, A. R., Evidence for both Wisconsin and Illinoian ages for the Tahoe glaciation, Sierra Nevada, California, *Geol. Soc. Am. Abstr. Prog., 16(6)*, 519, 1984.

Huber, N. K., Amount and timing of late Cenozoic uplift and tilt of the central Sierra Nevada, California - Evidence from the upper San Joaquin River, *U.S. Geol. Sur. Prof. Paper 1197*, 28 pp., 1981.

Janda, R. J., Pleistocene history and hydrology of the San Joaquin River, California, Ph.D. diss., 425 pp., University of California, Berkeley, Univ. Microfilms Inc., No. 68-5086, Ann Arbor, Michigan, 1966.

Mathieson, S. A., Pre- and post-Sangamon glacial history of a portion of Sierra and Plumas Counties, California, M.S. thesis, California State University, Hayward, 1981.

Marchand, D. E. and A. Allwardt, Late Cenozoic stratigraphic units, northeastern San Joaquin Valley, California, *U.S. Geol. Sur. Bull. 1470*, 1981.

Meierding, T. C., Late Pleistocene glacial equilibrium-line altitudes in the Colorado Front Range: A comparison of methods, *Quat. Res., 18*, 289-310, 1982.

Mezger, E., Pleistocene glaciation of the Cottonwood basin, southeastern Sierra Nevada, California, M.S. thesis, 149 pp., Univ. Southern California, Los Angeles, 1986.

Pierce, K. L., J. D. Obradovich, and I. Friedman, Obsidian hydration dating and correlation of the Bull Lake and Pinedale glaciations near West Yellowstone, Montana, *Geol. Soc. Am. Bull., 76*, 703-710, 1976.

Pierce, K. L. and W. E. Scott, Pleistocene episodes of alluvial-gravel deposition, southeastern Idaho, in *Cenozoic Geology of Idaho*, edited by Bill Bonnichsen and R. M. Breckenridge, *Idaho Bur. Mines and Geol. Bull., 26*, 685-702, 1982.

Porter, S. C., K. L. Pierce, and T. D. Hamilton, Late Wisconsin mountain glaciation in the western United States, in *Late-Quaternary Environments of the United States, Vol. 1, "The Late Pleistocene,"* edited by S. C. Porter, pp. 71-111, Univ. Minnesota Press, Minneapolis, 1983.

Shakleton, N. J. and N. D. Opdyke, Oxygen isotope and paleomagnetic stratigraphy of equatorial Pacific core V28-328: Oxygen isotope temperatures and ice volumes on a 105 year and 106 year scale, *Quat. Res., 3,* 39-55, 1973.

Sharp, R. P., Semiquantitative differentiation of glacial moraines near Convict Lake, Sierra Nevada, California, *J. Geol., 77,* 68-91, 1969.

Sharp, R. P., Pleistocene glaciation, Bridgeport Basin, California, *Geol. Soc. Am. Bull., 83,* 2233-2260, 1972.

Sharp, R. P. and J. H. Birman, Additions to the classical sequence of Pleistocene glaciations, Sierra Nevada, California, *Geol. Soc. Am. Bull., 83,* 2233-2260, 1963.

Smith, G. I., Subsurface stratigraphy and geochemistry of late Quaternary evaporites, Searles Lake, California, *U.S. Geol. Sur. Prof. Pap.., 1043,* 103 pp., 1979.

Smith, G. I., Paleohydrologic regimes in the southwestern Great Basin 0-3.2 m.y. ago compared with other long records of "global" climate, *Quat. Res., 22,* 1-17, 1984.

Wesling, J. R., Glacial chronology and soil development in Winsor Creek drainage basin, southernmost Sangre de Cristo Mountains, New Mexico, Ms. thesis, 165 pp., Univ. New Mexico, Albuquerque, 1988.